MEMORANDUM
RM-4989-TAB
AUGUST 1966

THE EFFECTS OF NUCLEAR WAR ON THE WEATHER AND CLIMATE

E. S. Batten

PREPARED FOR:
TECHNICAL ANALYSIS BRANCH
UNITED STATES ATOMIC ENERGY COMMISSION

The RAND Corporation
SANTA MONICA • CALIFORNIA

MEMORANDUM
RM-4989-TAB
AUGUST 1966

THE EFFECTS OF NUCLEAR WAR ON THE WEATHER AND CLIMATE

E. S. Batten

This research is sponsored by the San Francisco Operations Office of the U. S. Atomic Energy Commission under Contract AT(04-3)-414, Project Agreement No. 3. Views or conclusions contained in this Memorandum should not be interpreted as representing the official opinion or policy of the Atomic Energy Commission.

The RAND Corporation
1700 MAIN ST • SANTA MONICA • CALIFORNIA • 90406

PREFACE

The popular question, "Does the detonation of nuclear weapons affect the weather?" is here examined in a wider context, including the possible effects on climate of massive detonations such as might be expected in a major war. The author examines some of the most probable ways of affecting the weather and shows that our knowledge is yet insufficient to provide unambiguous answers.

This Memorandum represents one part of work being done by RAND for the United States Atomic Energy Commission, Division of Biology and Medicine, Technical Analysis Branch (TAB), on the biological and environmental consequences of nuclear war.

SUMMARY

The possibility that the energy, the debris, or the radioactivity of nuclear detonations in large number may affect the climate and the weather is explored. Because of the complexity and the lack of thorough understanding of the interdependent meteorological processes, it is impossible today to predict the consequences of artificial atmospheric stimuli. This paper explores ways that the by-products of a nuclear war may interfere with the dynamical, hydrological, and radiational processes in the atmosphere. The difficult problems of estimating quantitatively the weather and climatic changes are avoided except to emphasize the ambiguities of such estimates. The study indicates that the interference with the atmospheric processes in some cases can be sufficient to produce changes in them; however, it stresses that the nature, extent, and magnitude of the resulting anomalies in the weather and the climate are uncertain.

CONTENTS

I. INTRODUCTION

The testing of nuclear weapons was accompanied by considerable speculation about effects on the weather. Finding that the phenomena produced by the explosions were orders of magnitude smaller than similar, naturally occurring phenomena, most scientists concluded that the tests could create no measurable effect on the weather at distances greater than a few miles from the test site. However, the possibility of nuclear war remained, and the effort to evaluate the post-war environment increased. Changes in weather and climate were among the many problems considered. If a large number of weapons were to be detonated, the magnitude of the bomb-produced phenomena could approach the magnitude of natural phenomena. This seemed to eliminate the arguments that weather anomalies could not result from nuclear detonations. One study presented an argument that a nuclear war might produce an ice age (Stonier, 1963). Another presented a chronological table of possible meteorological events in the decade following a large-scale nuclear war (Ayres, 1964). While some such conjectures must be conceded possible, they should be tempered with the understanding that they can be neither proved nor disproved. This uncertainty stems from the lack of understanding of many complex and interdependent meteorological events. Nevertheless, because of the impact of weather and climate on man's activities, it is particularly important to give some consideration to possible changes during a post-war recovery period.

While it is not possible to predict with certainty the effects on the weather of such artificial stimuli as those resulting from nuclear

detonations, critical appraisals can point to the most probable sources of weather anomalies, and some may indicate at least the initial responses of the atmosphere. But the culmination of the effects remains unpredictable. Until further knowledge is gained of how man can intentionally or may accidentally interfere with the weather processes, we must be content with a less-than-satisfactory appraisal of effects of nuclear war on the atmosphere.

Before seeking sources of weather anomalies that may be produced by a nuclear war, we should arm ourselves with the general principles of atmospheric behavior. The next section will be a short outline of important atmospheric processes, particularly emphasizing areas where interference might succeed. With this information we can then examine the suggested effects of nuclear detonations on atmospheric processes to determine those most likely to produce changes in weather or climate. The atmosphere's initial response will subsequently be indicated for some cases, but pronouncements on the final or equilibrium state will be avoided because that state cannot be reliably described.

Rather than specify the possible meteorological consequences of nuclear war without knowledge of their likelihood, this paper will point out the probable sources of weather change and, in doing so, direct attention toward areas where improved understanding of the atmosphere will aid in the solution of this important problem.

II. SOME PRINCIPLES OF ATMOSPHERIC BEHAVIOR

This section attempts to provide the fundamental knowledge of atmospheric behavior needed to evaluate the credibility of several proposed theories of weather modification resulting from a nuclear war. But at the same time it is hoped that the discussion will provide a better appreciation of the complex interactions of several atmospheric processes. It should aid in the identification of stimuli most likely to produce weather changes but will also stress that, although we can estimate how changing a given parameter may modify a second parameter, the ultimate effects on the atmosphere as a whole are difficult to determine. It may even be that the ultimate consequences would tend to damp out the initial effect.

The atmospheric behavior is controlled by three distinct, yet interdependent, processes: the radiational, the dynamical, and the hydrological. The source of energy for the atmosphere is the radiation from the sun. Therefore, a description of the disposition of radiation reaching the atmosphere will provide a logical starting point.

The global average flux of solar radiation* reaching the top of the atmosphere is 0.5 cal cm^{-2} min^{-1} or 720 cal cm^{-2} day^{-1}. This radiation is dispersed by scattering, absorption, and reflection within the atmosphere and at the earth's surface. Earth and atmosphere too radiate energy, but at wavelengths longer than the sun's. Since, on the average, the earth and atmosphere are neither warming nor cooling, we assume the solar heat gain is exactly balanced by a terrestrial heat loss. The annual heat budget, averaged for the whole earth (London, 1957) is shown in Fig. 1.

*The global average flux is given by $S\pi a^2/4\pi a^2$, where S is the solar constant and a is the radius of the earth.

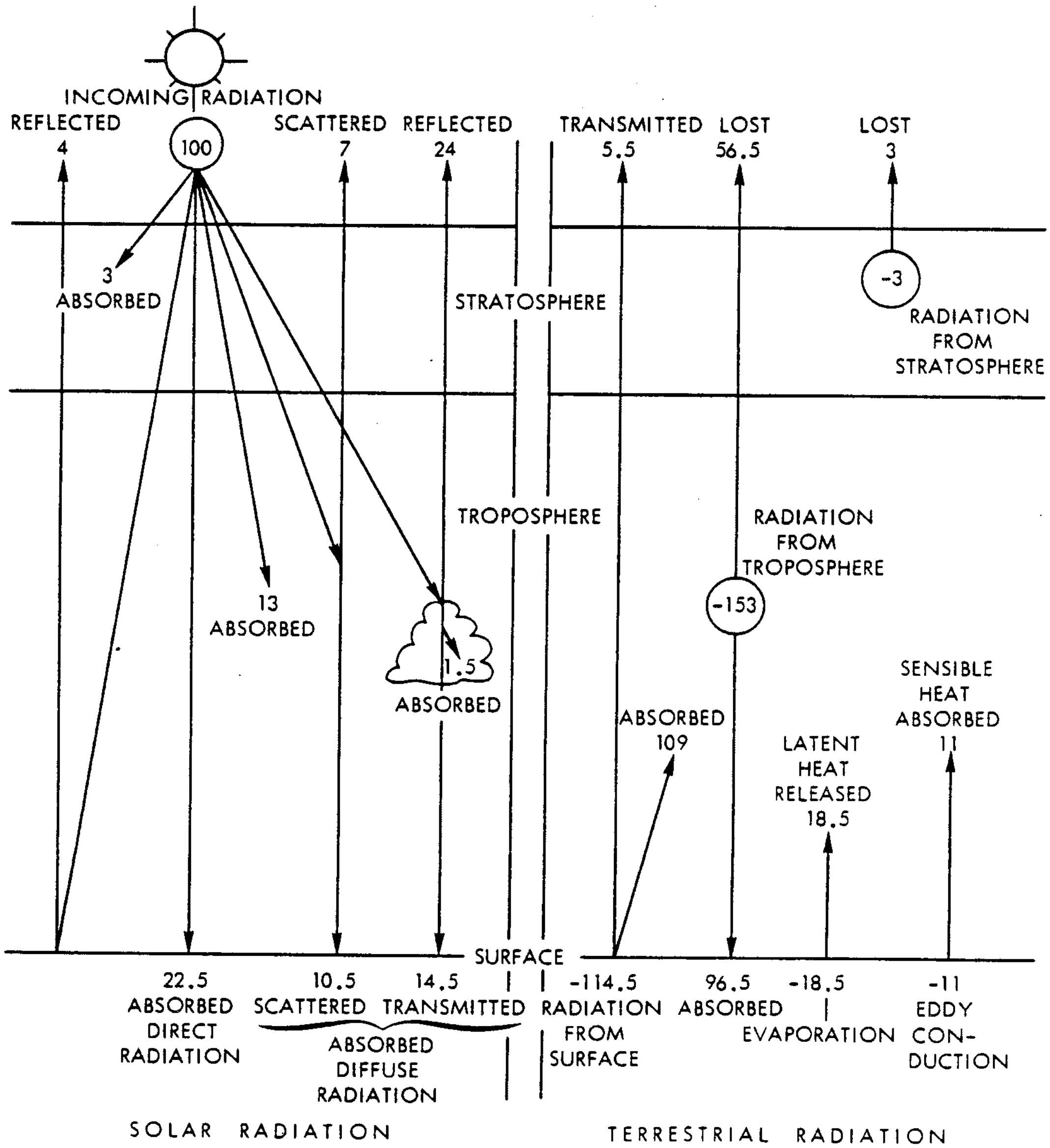

Fig. 1 — The annual heat budget averaged for the whole earth as given by London (1957).

In this figure the amount of radiation reaching the top of the atmosphere is taken as 100 units. To summarize the data in this figure, 35 units of solar radiation are returned to space by reflection and scattering, whereas the remaining 65 units are absorbed by the atmosphere and the ground. This gain of 65 units is balanced by a loss of an equal number of units, partly by a transmission through the atmosphere of radiation from the ground, but mostly by radiation from the troposphere and stratosphere. It should be noted that radiation is not the only source of heat for the atmosphere. The atmosphere also gains heat by turbulent transfer from the surface and through evaporation with eventual condensation in the atmosphere. The net value of these nonradiational processes appears insignificant when the global heat budget is considered. However, an investigation of the heat budget for various latitudes will adequately show their importance. This point will be considered next.

The net gain of radiative heat by the earth and atmosphere for various latitudes is given in Fig. 2. These results by Budyko (1956) show a surplus of radiation at low latitudes and a deficit at high latitudes. If such a condition were allowed to continue without a redistribution of heat, clearly the tropical regions would become progressively warmer while the poles would become colder. The heat is redistributed by ocean currents and by horizontal fluxes of sensible and latent heat in the atmosphere. The mechanisms of heat transport in the atmosphere are significant, since they are an important link between the radiative, hydrological, and dynamical processes. The simplest mechanism providing a north—south heat transport from the

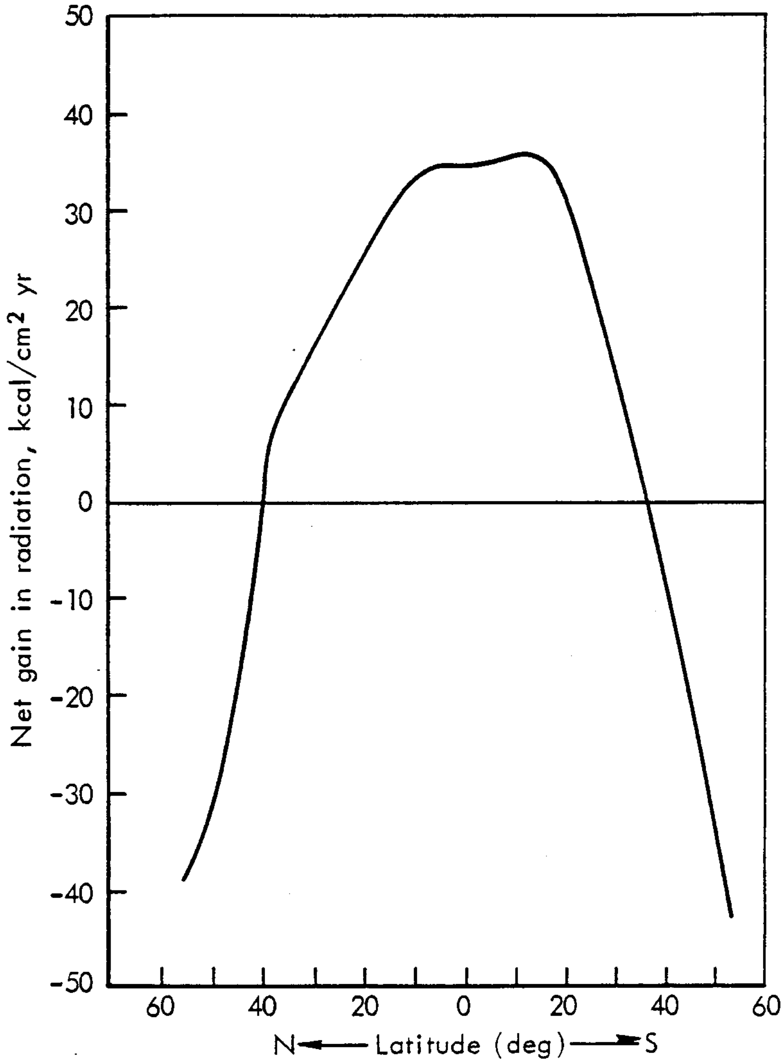

Fig. 2 — Net gain of radiation by the earth and atmosphere. (After Budyko.)

equator to the poles is a direct meridional circulation often referred to as the Hadley regime. However, the meridional flow necessary to produce the required transport of heat by itself, would also produce a dynamically unstable east—west flow in the atmosphere. Instead, the atmospheric flow at mid latitudes departs markedly from a symmetrical circumpolar vortex and actually consists of quasi-horizontal waves. The air in these disturbances moves from low levels and low latitudes upwards and poleward while air from the upper levels and higher latitudes moves downward and equatorward. Moisture condenses in the warm air moving poleward, forming clouds and precipitation, thus providing the required transport of sensible and latent heat. The waves just described are the frontal systems responsible for weather changes at mid latitudes. At low latitudes the Hadley type of circulation is the primary mechanism for transporting heat and moisture. The transport of latent and sensible heat by the atmosphere and the transport of heat by the oceans are shown in Fig. 3. In this figure the positive values show a gain of heat by the earth and by the atmosphere at any given latitude. The heat gained or lost by these processes exactly balances the net loss or gain shown for the same latitude in Fig. 2. The radiational processes supply heat to the earth and atmosphere either directly through absorption or indirectly through the hydrological cycle of evaportation, condensation, and precipitation. It is the dynamical processes resulting from heat imbalances, then, that cause heat and moisture to be redistributed within the atmosphere.

The importance of cloud formation should not be overlooked. In Fig. 1 it will be noted that 24 per cent of the incoming solar radiation is reflected by clouds. Increasing or decreasing clouds can significantly

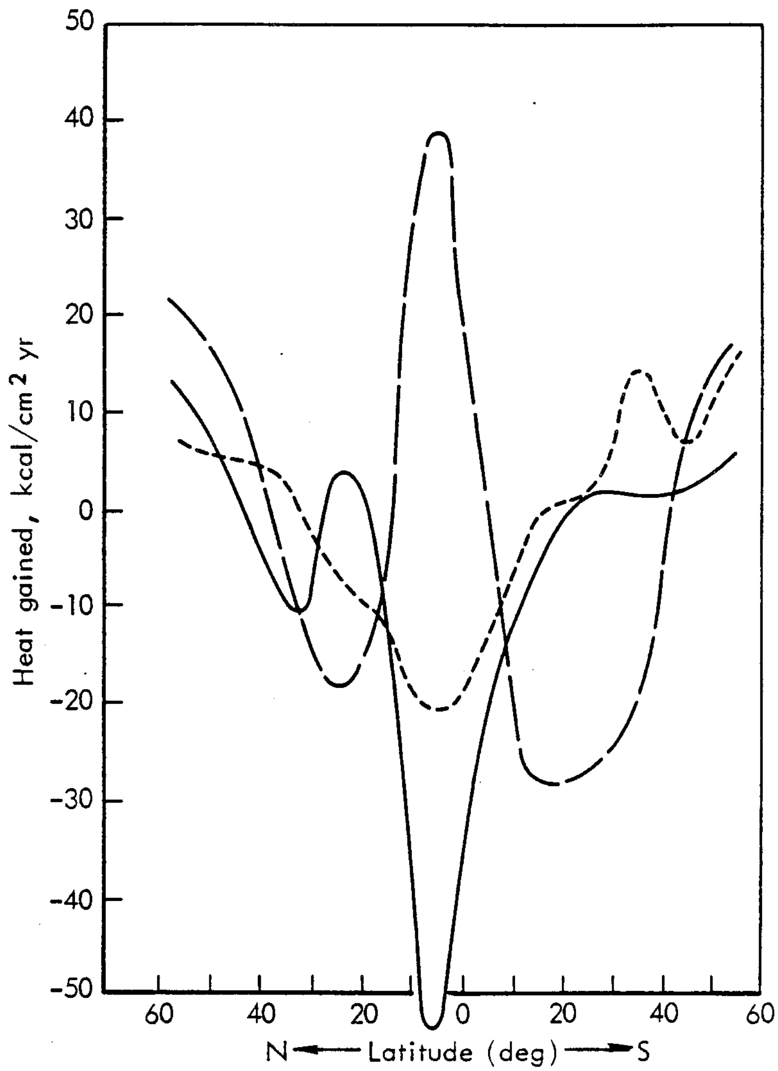

Fig. 3 — Heat gained by the earth—atmosphere system from sensible heat transports in the atmosphere (———), transports in the ocean (----), and latent heat transports in the atmosphere (— —). (After Budyko.)

affect the amount of energy available to the atmosphere and thus influence the temperature (Ohring and Mariano, 1964).

In theory, the weather could be modified by disturbing any one of the three processes discussed above. However, it should be clear that the ultimate effects on the weather are not confined to the disturbed process; the other processes will also be affected to some extent.

These are the basic principles of atmospheric behavior. Some pertinent details will be discussed in the next section, in which suggested effects of nuclear war on the weather are examined.

III. SOME SUGGESTED EFFECTS OF NUCLEAR DETONATIONS ON THE WEATHER

The energy, the debris, and the radioactivity produced by nuclear detonations have been suggested as possible causes of weather anomalies. This section will explore how these by-products of a nuclear war may interfere with the dynamical, hydrological, and radiational processes in the atmosphere and determine whether the interference appears sufficient to produce significant changes in them.

Effects on Dynamical Processes

The awesome event of a nuclear detonation suggests to many that the energy released may produce changes in the dynamical behavior of the atmosphere. Now if we ask how the energy of a 20-MT bomb compares with the 2×10^{21} cal absorbed daily by earth and atmosphere together, we find that one such bomb exploded approximately every second of the day would not quite match it.

A single feature of the atmosphere, however, seems at first glance to be more nearly comparable in potency to the bomb. A storm system covering 15×10^5 km^2 has an estimated instantaneous value of kinetic energy equivalent to 4×10^{16} cal, or just twice that of the 20-MT bomb. However, the storm requires a constant supply of energy to counteract frictional dissipation. If the rate of dissipation were 1 cal m^{-2} sec^{-1} the storm would last about 8 hours. To maintain the storm against this dissipation for the normal 5- to 6-day lifetime requires energy the equivalent of that from a 90-KT bomb every minute or a total of 650 to 780 megatons during the lifetime.

But the 1-minute duration of one nuclear explosion and the few square kilometers it affects are not appropriate scales of energy

release for maintaining a process of the duration and scope of a storm system.

The effects of a rapid release of explosive energy over a small area are well known. High-frequency pressure waves of the acoustical--gravitational type comparable to those produced by the 58-MT Soviet explosion in October 1961 also followed the eruption of Krakatoa on August 26, 1883 (Wexler, 1962). Waves of this type radiate from the source of the explosion much as do the waves from a rock dropped in a pond. On a spherical earth, however, they travel to the antipodes where they are reflected toward the source. After the Krakatoa and Soviet detonations, the waves traveled several times around the earth. It is doubtful that waves of this type could have any effect on the storm-producing systems of the atmosphere. Even if all the energy in such waves, produced by detonations totaling 10,000 MT, were dissipated uniformly around the earth within the lowest 1 km, the resultant heating of that part of the atmosphere would be less than 0.1°C.

Therefore, it seems that the energy released by nuclear detonations could not supply the conditions required for storm formation. It has been suggested, however, that a nuclear detonation may disturb an already dynamically unstable flow and thus "trigger" a storm by releasing pent-up natural forces. This may have been so at Bikini in July 1946. It has been theorized that the underwater detonation ejected vast quantities of water into the atmosphere, and caused a potentially unstable air mass to rise, thus producing clouds and about 30 minutes of rain. It is likely, though, that precipitation would have occurred without the stimulus of the bomb. Granting that a large number of detonations within a small area could trigger a storm, one might then ask whether

the storm could continue to affect subsequent weather patterns. To produce long-lasting effects over an extended area would require the modification of conditions external to the storm either by detonations or possibly by the storm itself. It does not seem reasonable to assume that the general intensity or frequency of storms over an extensive area and for periods as long as a season could be affected, since both the intensity and frequency of future storms are not necessarily consequent upon the initial disturbance. Rather, they are affected by the large-scale patterns of solar heating in the atmosphere and depend upon the dynamical stability criteria inherent in the atmospheric flow. Any lasting effect of the artificially induced storm would be confined to the immediate locality and would be sustained only as long as some local imbalances were sustained. An imbalance might be, for example, a storm-induced "feedback" mechanism, creating only minor changes in storm tracks. For example, Namias (1962) has suggested that, in 1960, an anomalous snow cover at low latitudes may have caused a local atmospheric loss of heat, thus creating conditions favorable for storm generation at lower-than-normal latitudes. It is not known whether the anomalous snow cover could have resulted from a single abnormal event or whether it should be attributed to a larger-scale change in the normal atmospheric circulation. However, the snow cover seemed to play an important role in maintaining itself and in modifying the normal storm track.

In summary, it is not probable that the weather would be affected by the addition to the atmosphere of energy released by bombs. Although it may be argued that a storm can be triggered, the effects would be short-lived unless the storm could produce feedback to the atmosphere. Significant changing of the weather by the energy released through

nuclear detonations seems unlikely, because those very conditions necessary to permit artificial weather changes would probably create similar effects in the absence of the artificial stimuli.

Effects on Hydrological Processes

The large quantities of dust produced by a surface detonation, the ions produced by all nuclear detonations, and the modification of the atmosphere's electrical properties resulting from the increased ionization have been suggested as stimuli that may increase cloudiness and precipitation. Some of these suggestions are the result of misconceptions of the cloud- and precipitation-forming mechanisms. In other cases it must be admitted that too little is known about the microphysics of clouds and precipitation processes to state unequivocally what these stimuli might do. A discussion of the theories of cloud and precipitation formation will be helpful in sorting out the suggested effects of nuclear detonation products on these processes.*

It has long been observed that in pure air (air without foreign particles and ions) the relative humidity can reach about 800 per cent. Since humidities in the atmosphere rarely exceed 100.1 per cent, we must have particles called condensation nuclei to initiate cloud formation at such a realistic humidity. The nuclei in the atmosphere can be classed as nonsoluble wettable, nonsoluble hydrophobic, and soluble hygroscopic. The nonsoluble hydrophobic nuclei can be disregarded for our purposes because, as the name implies, they tend to repel water; to start condensation requires humidities that do not occur in the atmosphere. For nonsoluble wettable nuclei, the required degree of

---

*A more comprehensive discussion of cloud and rain formation can be found in Clouds, Rain, and Rainmaking by B. J. Mason (1962).

supersaturation depends primarily on their radii, r,

$$\ln S = \frac{2\sigma M}{\rho RT} \frac{1}{r} \tag{1}$$

where S is the saturation ratio (or humidity if expressed as percentage), σ is the surface tension of the water surrounding the nuclei, ρ is the density of water, R is the universal gas constant, M is the molecular weight of water, and T is the temperature. Some values from this equation are given in Table 1.

Table 1

VALUES OF THE HUMIDITY REQUIRED
TO INITIATE CONDENSATION ON PARTICLES OF RADIUS r
(The values used in the computation are: M = 18; $R = 8.314 \times 10^7$ erg mol^{-1} $°K^{-1}$; $\sigma = 72.75$ dyne cm^{-1}; T = 293°K; $\rho = 1$ g cm^{-3})

Nucleus Radius, r (μ)	Humidity (per cent)
10^{-3}	293.0
10^{-2}	111.4
10^{-1}	101.1
10^{0}	100.1
10^{1}	100.01

For humidities occurring in the atmosphere (<100.1 per cent), the nonsoluble wettable nuclei must be larger than 1μ, (10^{-4} cm) to start condensation. A charge on the nonsoluble particle does decrease the degree of supersaturation required for condensation, but the charge effect is too small to reduce significantly the required size of the nucleus. For example, including the effect of an electrical charge, Eq. (1) becomes

$$\ln S = \frac{2\sigma M}{\rho RT}\frac{1}{r} - \frac{q^2 M}{8\pi\epsilon\rho RT}\frac{1}{r^4} \qquad (2)$$

where q is the electrical charge on the particle, ϵ is the dielectric constant for air (taken equal to 1), and the other symbols are the same as in Eq. (1). For very small particles -- aggregates of a few molecules -- the second term on the right of Eq. (2) dominates. But even in the presence of a charge, humidities much greater than 100 per cent are needed for particle growth. A particle of radius $6 \times 10^{-4}\mu$ with a charge of one electron ($q = 4.8 \times 10^{-10}$ e.s.u.) requires a humidity of 356 per cent. For particles near 1μ radius the effect of the charge is negligible for reasonable values of q. For a smaller particle, say, $10^{-2}\mu$, a charge equivalent to more than 100 electrons is needed to start condensation at humidities found in the atmosphere. According to Greenfield (1955), it is unlikely that debris from nuclear detonations will have a net charge. Therefore, ions and radioactive particles will not be preferred as nuclei for condensation. Although the number of large nonsoluble nuclei in the atmosphere will increase, they will have to compete with the natural nuclei for available moisture. Of the natural nuclei in the atmosphere soluble hygroscopic nuclei are the most effective, for condensation on soluble hygroscopic nuclei can start at relative humidities below 100 per cent because the presence of solute in the cloud droplet reduces the equilibrium vapor pressures of the surface. For humidities near 100 per cent, it can be shown that

$$S = 1 + \frac{2\sigma M}{\rho RT}\frac{1}{r} - \frac{8.6m}{M_1}\frac{1}{r^3} \qquad (3)$$

where m is the mass of the solute and M_1 is the molecular weight of the solute. Common types of hygroscopic nuclei are sulfuric acid and salts such as sodium chloride. The combustion process furnishes finely divided carbon, and the more important sulfur dioxide which, through the oxidizing action of the sunlight, becomes sulfuric acid. Using Eq. (3) it can be shown that a sodium chloride crystal of radius $1.03 \times 10^{-1}\mu$ and mass of 10^{-14} gm will grow to a droplet of one-micron radius if the humidity is 100 per cent. If the humidity exceeds the critical value (in this case 100.042 per cent), the droplet can grow without limit. A nonsoluble nucleus the same size as the salt crystal would require a humidity of 101.1 per cent to initiate condensation, a value of humidity that exceeds the critical value for unlimited growth on the salt crystal. In nature, many droplets are competing for moisture at the same time. If water vapor is not available at a rate greater than the growth rate of all the droplets, the humidity decreases and the growth is checked. Clearly these conditions for condensation favor the hygroscopic nuclei. Because there seems to be an abundance of hygroscopic nuclei, only the largest nonsoluble nuclei will be active.

It is hard to predict the effect that an increased number of large nonsoluble nuclei might have on cloud formation. Nature gives some hints. For example, concentrations of nuclei vary from 10^5 particles per cm^3 near industrial cities to 10^3 particles per cm^3 over ocean areas. This large difference in concentration does not seem to affect the development of clouds. However, there are some indications that clouds over ocean areas contain fewer, but larger, droplets ($10/cm^3$

of radius ~25μ) than the clouds formed over continents ($10^3/cm^3$ of radius 2 - 5μ). Therefore, by increasing the number of large nonsoluble nuclei the most likely effect, if any, would be an increased number of smaller cloud droplets but no significant change in cloudiness. Observation seems to show that there are always sufficient nuclei to form clouds. It is not justifiable, therefore, to assume that the water content of the clouds would increase, since the availability of moisture for condensation is unrelated to the number of nuclei present. The possible change in cloud-droplet size-distribution might have some effect on the efficiency of precipitation formation for reasons to be considered next.

Describing the mechanisms by which cloud droplets can grow to sizes large enough to fall as some form of precipitation is a major problem of cloud physics. The two commonly proposed processes -- the Bergeron ice-crystal process and the coalescence process -- may be affected by a large number of nuclear detonations. Briefly, the ice-crystal theory states that ice crystals, when surrounded by supercooled water droplets, will grow at the expense of the water droplets, since vapor pressure is less over ice than over water. In other words, when the air is saturated with respect to water it is supersaturated with respect to ice and condensation on the ice crystal will result. Water droplets are observed in clouds often at temperatures as low as -20°C and sometimes -35°C; thus if ice crystals are introduced into the cloud, the stage is set for creation of large drops. However, as in the case of condensation, the initiation of freezing requires a suitable nucleus. The most active ice nuclei are nonsoluble wettable particles of dust

and soil about 1μ in radius. Silicate minerals of clay and mica were found to produce one ice crystal for every 10,000 particles in supercooled clouds at temperatures of -15°C or higher. The most abundant mineral is kaolinite, which is active at a temperature of -9°C. The prevalence of liquid-water clouds down to -15°C suggests that efficient ice nuclei are rare, at least at the altitudes where ice crystals "seed" the supercooled clouds. It is not unreasonable to suspect that the efficiency of the Bergeron process might increase if the large amounts of debris produced by surface detonations can act as ice nuclei.

The coalescence theory states that large cloud droplets falling relative to smaller ones grow by collecting the smaller droplets. An extension of this theory indicates an increase of collection-efficiency if the cloud droplets are falling through an electrical field or if they contain a net charge. Following a nuclear explosion, the ionization of the air increases, thus increasing the conductivity of the air and decreasing the potential gradient in the atmosphere. An increasing droplet charge and a decreasing potential gradient have opposite effects on the collection efficiency; therefore, the net effect on precipitation is uncertain.

One may conclude, first, that the number of water clouds will probably be unchanged as a result of an increase in the number of nonsoluble nuclei, but the mean radius of the cloud droplets may decrease and the concentration of droplets may increase. Decreasing the mean radius may slow droplet growth and inhibit precipitation. Second, the nonsoluble particles may act as ice nuclei if they are dispersed high enough in the troposphere, and thus increase ice-cloud formation.

The increased number of ice crystals, under proper conditions may enhance the precipitation process. Third, the net effect on precipitation of the increased ionization in the atmosphere is indeterminable at present.

Clearly the effect of bomb debris on precipitation is unknown. But if there is an effect, how long could one assume it to last? If the debris were confined to the troposphere, the precipitation changes could last only a few months, since rainfall is a very efficient scavenger of atmospheric contamination. However, a large number of small particles would probably spread throughout the stratosphere, slowly settling out in two or three years. If the stratospheric debris is successful in modifying the precipitation, the long-term effects on the hydrological cycle and heat balance of the atmosphere must be determined before predictions of weather and climate change can be made. Any long-term increases or decreases of precipitation must be accompanied by an acceleration or deceleration of the hydrological cycle. Adjustments of the heat budget should follow, since the transport of latent heat would probably be affected. As an example, consider the consequences of changing the water balance north of 25° latitude. According to Starr, <u>et al</u>. (1957), the average amount of water vapor transported across 25°N by the atmospheric motion is near zero (Fig. 4). Therefore, the average annual precipitation north of 25° must be balanced by the evaporation in this region. The precipitable water-vapor content of the atmosphere north of 25° averages 1.53 gm/cm^2 (Fig. 5). If the annual average precipitation for this region (72.6 cm) were to increase by 10 per cent, all other processes remaining the same, the supply of water vapor would be depleted in less

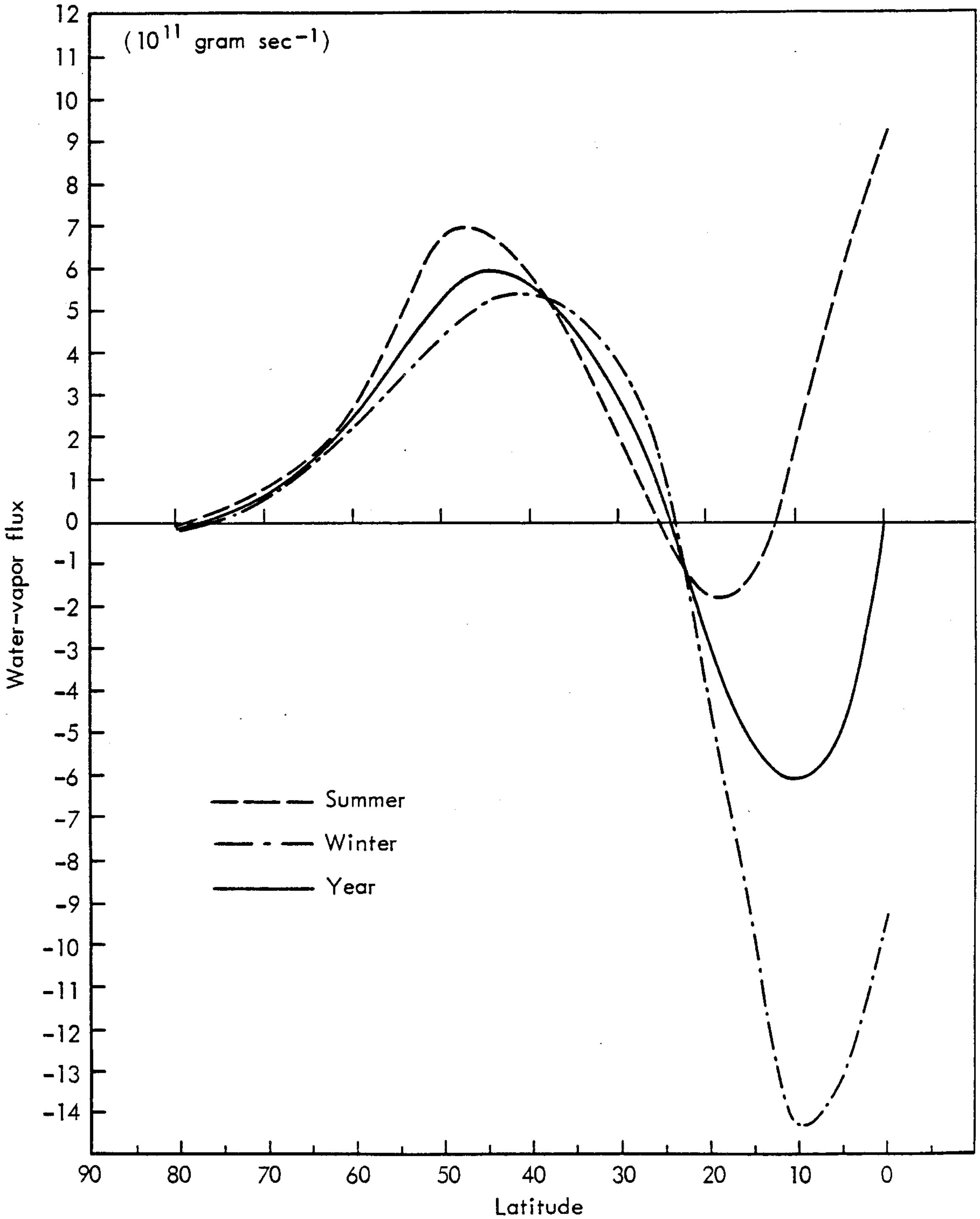

Fig. 4 — The meridional distribution of zonally averaged water-vapor flux in the atmosphere computed from atmospheric data. (After Starr, et al., 1957.)

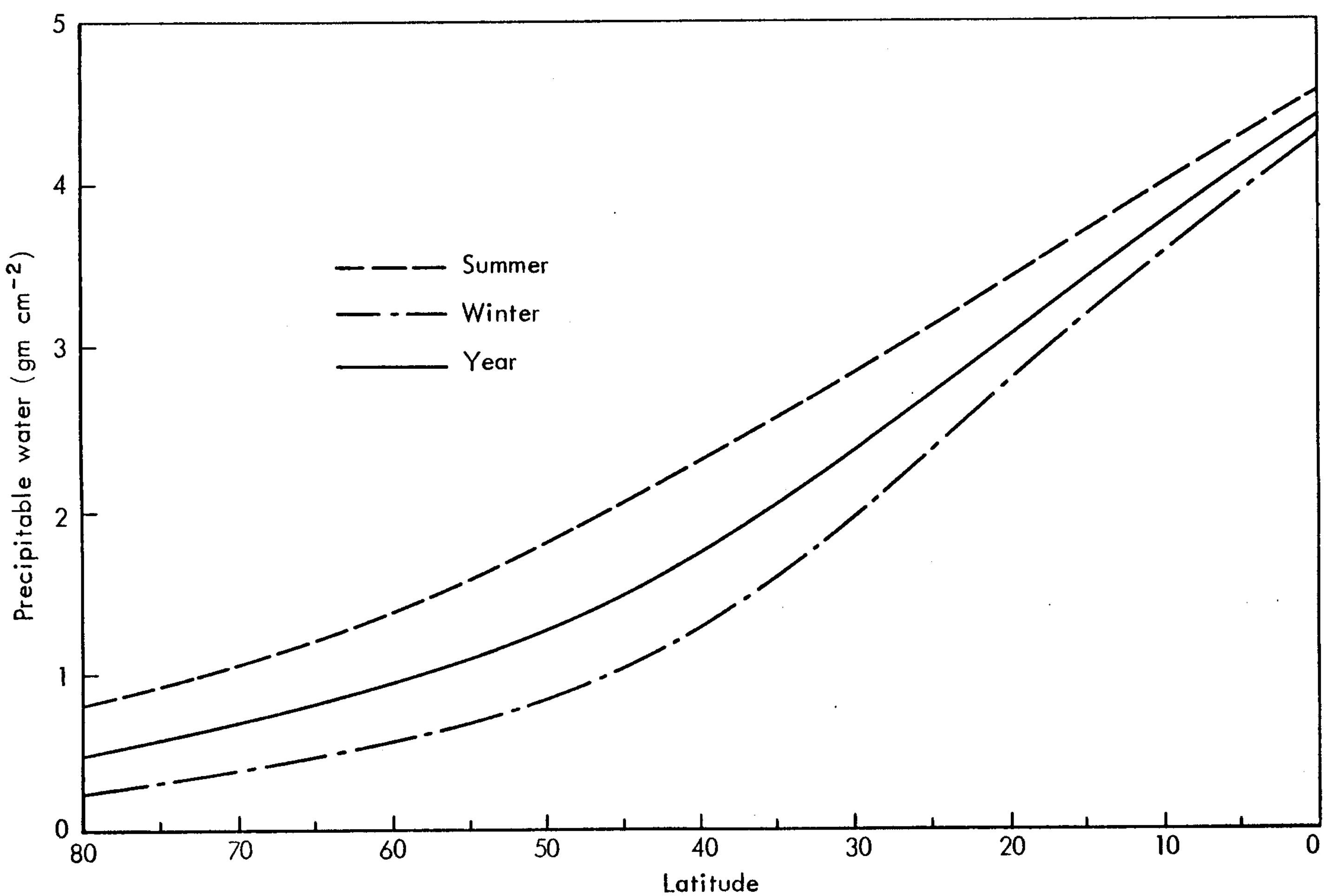

Fig. 5 — The meridional distribution of zonally averaged precipitable water vapor computed from atmospheric data. (After Starr, et al., 1957.)

than three months. Obviously, other atmospheric changes must occur if an increase of precipitation is to be maintained for long periods. Either increased evaporation or a different circulation in low latitudes, resulting in a moisture transport across 25°N, could provide the necessary moisture. Both would have far-reaching consequences on the weather patterns.

However, it is not reasonable to assume that the circulation would respond immediately to the increased precipitation without intervening changes in the heat budget. The evaporation would be expected to increase, following an initial "drying out" trend in the atmosphere, since evaporation is in part a function of the vertical gradients of moisture near the surface. Increasing the evaporation would ultimately result in a modification of the heat balance at the surface. Within any latitude belt, the average net gain of heat through radiational processes at the surface is equal to the sum of heat losses by evaporation and by turbulent transfer, and the heat loss or gain through horizontal transport in the oceans. Changes in the ocean transport would probably be small, at least initially, and since the mechanisms of vertical sensible-heat transfer and latent-heat transfer are closely related, it is unlikely that the increasing heat loss by evaporation will be balanced by a decrease in turbulent heat transfer alone. The most probable source of the heat needed for evaporation is the increase in the net gain of radiation at the surface either by an increased insolation at the surface (implying a decrease in atmospheric albedo and, therefore, in clouds) or by decreased longwave radiation from the surface (implying a lower surface temperature). Only the latter seems reasonable, assuming an increase in precipitation. These modifications of

the heat budget could lead to changes in the atmospheric circulation, but it is difficult to determine the nature of the changes. If all the processes proceed at a suitable rate, the circulation could change the whole character of its behavior and remain changed until the disturbing influence was removed. On the other hand, if any one of the processes failed to change in the proper proportion or at the proper rate, it is possible that the initial tendency toward increased precipitation would be inhibited. For example, if the rate of evaporation failed to increase, or if the moisture transport process failed to intensify, the rate of precipitation could not remain at the higher level without "drying out" all or part of the atmosphere north of 25° latitude.

Effects on Radiational Processes

A more direct method of changing atmospheric processes would be to interfere with the incoming radiation. Both stratospheric dust and more frequent or thicker ice clouds can modify the incoming solar energy available to the atmosphere. This idea is not new. Humphreys (1913 and 1940) proposed that the large quantities of stratospheric dust produced during periods of extensive volcanic activity could be used to explain the ice ages. Measurements indicated that direct solar radiation decreased by as much as 20 per cent following such large eruptions as Krakatoa in 1883 and Katmai in 1912. Humphreys argued that this decrease of direct solar radiation, if maintained for long periods, would ultimately lower the mean surface temperature. Although there are some oversimplifications in the theory leading to this conclusion, it must be accepted that a decrease in the direct solar radiation by as much as 20 per cent could significantly modify the radiation balance of the earth and atmosphere.

Estimates of the amount of stratospheric dust produced by Krakatoa or Katmai are not precise. Wexler (1951b and 1952) quoted a value of 13 mi^3 ($\sim$ 54 km^3) for the amount of debris thrown into the atmosphere by the eruption of Krakatoa. Of this amount he assumed that 1/3 consisted of particles small enough to remain suspended in the stratosphere for long periods. However, estimates of the size and concentration of the ejected dust, made from studies of the atmospheric optical phenomena created by the dust, suggested much smaller quantities. For example, Humphreys estimated the diameter of the particles to be 1.85μ and the number of particles above a 1-cm^2 area to be 3.4 x 10^5, while Deirmendjian (1965)* estimated 1 to 1.2μ and 10^7, respectively. Assuming an even distribution over the earth, the estimated total volume of dust would be 5.7 x 10^{-3} km^3 (Humphreys) or from 2.7 x 10^{-2} to 4.6 x 10^{-2} km^3 (Deirmendjian). These later estimates are probably more valid than the much larger estimates of Wexler. Since the Krakatoa debris had a known effect on the direct solar radiation, the volume of 0.5 x 10^{-2} to 5 x 10^{-2} km^3 will be considered the criterion with which to gage the effectiveness of debris from a nuclear war.

Surface or near-surface detonations in the megaton range are required to produce stratospheric dust layers. The cloud of debris from submegaton detonations will not penetrate into the stratosphere, and high-altitude detonations cannot produce significant quantities of dust. For surface detonations, it is estimated that the cloud contains on the order of one ton of debris for each ton of explosive force. Thus the cloud from a one-megaton detonation will contain 10^6 tons of debris. Of this material, only a small fraction of the particles will be fine enough

---

*Private communication.

($\leq 2\mu$ in diameter) to remain in the stratosphere for long. It is estimated from measurements of the particle-size distribution in a nuclear cloud that 0.1 to 1 per cent of the material consists of particles less than 2μ (see Appendix A). Each MT detonation would, therefore, add between 10^3 and 10^4 tons of debris to the stratosphere. This is equal to a volume of 0.4×10^{-6} to 4×10^{-6} km^3, assuming a density of 2.5 tons per m^3. Therefore, detonations totaling 10,000 MT would produce between 0.4×10^{-2} and 4×10^{-2} km^3, a volume equivalent to the estimated amount of stratospheric debris after Krakatoa.

The residence time of stratospheric debris depends on the altitude and latitude of injection. Studies of radioactive debris (Staley, 1960 and Friend, 1961), have indicated residence half-times ranging from a few months to a few years. The largest values have been attributed to injections at high altitudes in the tropics. For example, Staley suggests an 18-month residence half-time for material injected into the lower stratosphere (about 20 km) in the tropics and a 9-month residence half-time at similar altitudes but higher latitudes. The mean cloud height of a 1-MT surface detonation is approximately 22 km and that of a 20-MT detonation approximately 34 km. Thus, debris from detonations in the megaton range will remain in the stratosphere for at least 2 to 3 years and possibly longer.

The suggested modifications of the weather resulting from volcanic debris in the stratosphere -- a lower mean surface temperature and an increased north—south temperature gradient -- are not indisputably supported by observation. A close examination of Humphreys' data will show, for example, that in many cases the eruptions have occurred at or near the bottom of a cooling trend and not before it, as one might expect

from a cause-and-effect relationship. In addition, the statistical significance of a 0.1°F (annual) and 0.16°F (winter) decrease of the mean hemispheric temperature for the five years following an eruption (Mitchell, 1961) must be considered carefully. Although the observational evidence does not clearly demonstrate a relationship between a stratospheric dust cloud and weather anomalies, it seems reasonable to assume that adjustments to the radiation budget will take place. Thus, a further look at the "vulcanism" theory is in order to determine the nature of these adjustments.

The vulcanism theory as presented by Humphreys assumes that the dust cloud will reflect rather than scatter solar radiation, since the dust particles are large compared to the wavelength of the solar radiation. On the other hand, the cloud will have very little effect on the terrestrial radiation, since the particles are small compared to the wavelength. It is further assumed that the small amount of terrestrial radiation scattered will be equally dispersed in the forward and backward directions. Neglecting absorption of radiation in the cloud, it is concluded in the theory that the loss of solar radiation by reflection from the cloud far exceeds the amount of terrestrial radiation trapped by the cloud. This condition, it is argued, should result in a net cooling of the earth and increased storminess.

The theory briefly described above was first presented by Humphreys in 1913 and propagated in subsequent printings of his classical textbook *Physics of the Air*. It does not, however, make proper use of scattering theory as it is known today. A proper use must include results of a theory for multiple scattering by absorbing particles. At the present time, results are available only for the problem of *single* scattering by

absorbing spheres. A description of these can be found in van de Hulst's (1957) book on the subject. Only a brief discussion will be given here in order to point out some of the difficulties in the vulcanism theory and to emphasize other possible effects of a stratospheric dust layer.

If radiation of wavelength λ is incident on a particle of radius r, the nature and amount of scattered and absorbed radiation can be shown to be a function of a parameter $x = 2\pi r/\lambda$ and the refractive index, $m = n - in'$, of the particles. The scattering equations can be simplified with certain limiting assumptions. For example, by assuming that x is small and the product $x(m - 1)$ is small, the solution to the complete equations reduces to that given by the Rayleigh scattering theory. This theory predicts that the scattered intensity is proportional to λ^{-4} and that the intensities in the forward and backward directions are equal. On the other hand, if both $(m - 1)$ and $x(m - 1)$ are small, the complete equations reduce to a form given by the Rayleigh—Gans theory, which predicts that the ratio of forward to back scattering increases as x increases. To evaluate Humphreys' vulcanism theory, we assume that $(m - 1)$ and $x(m - 1)$ are large. In the limiting case, where m approaches infinity, the equations reduce to a form for totally reflecting spheres. The solution shows for small x, particles small compared to the wavelength $(x < 1.38)$ that the scattering is predominantly backward. However, for large x $(x > 1.38)$ the scattering, as in the Rayleigh—Gans theory, has the usual preponderance in the forward direction. The apparent paradox of a "totally" reflecting sphere scattering large amounts of radiation in the forward direction is a result of diffraction of radiation by the particle. As x increases, that is, as the ratio of particle radius to wavelength increases, the amount of diffraction also increases. The

radius of the particles in the Krakatoa dust cloud has been estimated at 0.5 to 1.0μ. Therefore, for solar radiation ($\lambda \sim 0.5\mu$), x is between 6 and 12. For terrestrial radiation ($\lambda \sim 10\mu$), x is between 0.3 and 0.6. The scattering theory predicts, therefore, that although a larger amount of solar radiation will be scattered (x is large), most of this radiation will be scattered in the forward direction. The terrestrial radiation, however, will be mostly scattered backward. Without a proper solution to the scattering problem, the net effect of the cloud cannot be determined. Multiple scattering may be an additional complication if the concentration of particles is great enough.

Any absorption of solar or terrestrial radiation by the dust cloud will modify the scattering effect on the heat balance. The radiation absorbed will be reradiated partly to space and partly to the earth. Whether absorption by the cloud will tend to produce a warming or a cooling will depend upon the relative amounts of solar and terrestrial radiation absorbed by the dust layer and upon the efficiency of the dust particles in reradiating the heat.

The modification of the heat balance can be determined only by first applying a radiative transfer theory including the effects of multiple scattering and absorption. The results must also account for the changes in the heat balance that can be expected to follow the temporal changes of the number (and size) distribution of particles in the dust layer.

While a solution to this complex problem does not exist at this time, we may look at the possible initial trends in the heat balance by assuming various combinations of absorption and scattering effects in the dust layer. Emphasis must be placed on "initial trends" since

the subsequent adjustments of the dynamical processes and the hydrological cycle cannot be precisely determined. As we have seen, the three processes -- radiational, dynamical, and hydrological -- are all important in determining the equilibrium values of the heat balance.

Equations for the effects of a dust cloud on the mean annual heat budget for the earth and atmosphere have been derived in Appendix B, where it is shown that the conditions for warming or cooling the earth's surface or the atmosphere are given by inequalities in the form:

$$\frac{(a' + s')}{2} = \beta \gtreqless \Phi\ (a, s, f, \rho, \gamma, \alpha, \tau, \delta) \begin{cases} \text{warming} \\ \text{no change} \\ \text{cooling} \end{cases} \qquad (4)$$

where β is the fraction of terrestrial radiation trapped by the cloud, and a' and s' are the fractions of the longwave radiation absorbed and scattered by the dust layer. The values of Φ for the earth's surface are different from the values for the atmosphere, but in both cases they are a function of the fraction of solar radiation absorbed (a) and scattered (s) by the layer, the degree of forward scattering (f), the albedo of the earth and atmosphere (ρ), the fraction of solar radiation absorbed by the earth (γ) and atmosphere (α), the transmissivity of the atmosphere to longwave radiation (τ), and the fraction of the radiation from the atmosphere returned to the earth (δ). Complete graphs of the function for the earth's surface and the atmosphere can be found in Appendix B. Since the expected value of the attenuation of solar radiation is approximately 20 per cent, only the portion of the graphs given in Figs. 6 and 7 need be considered. These figures

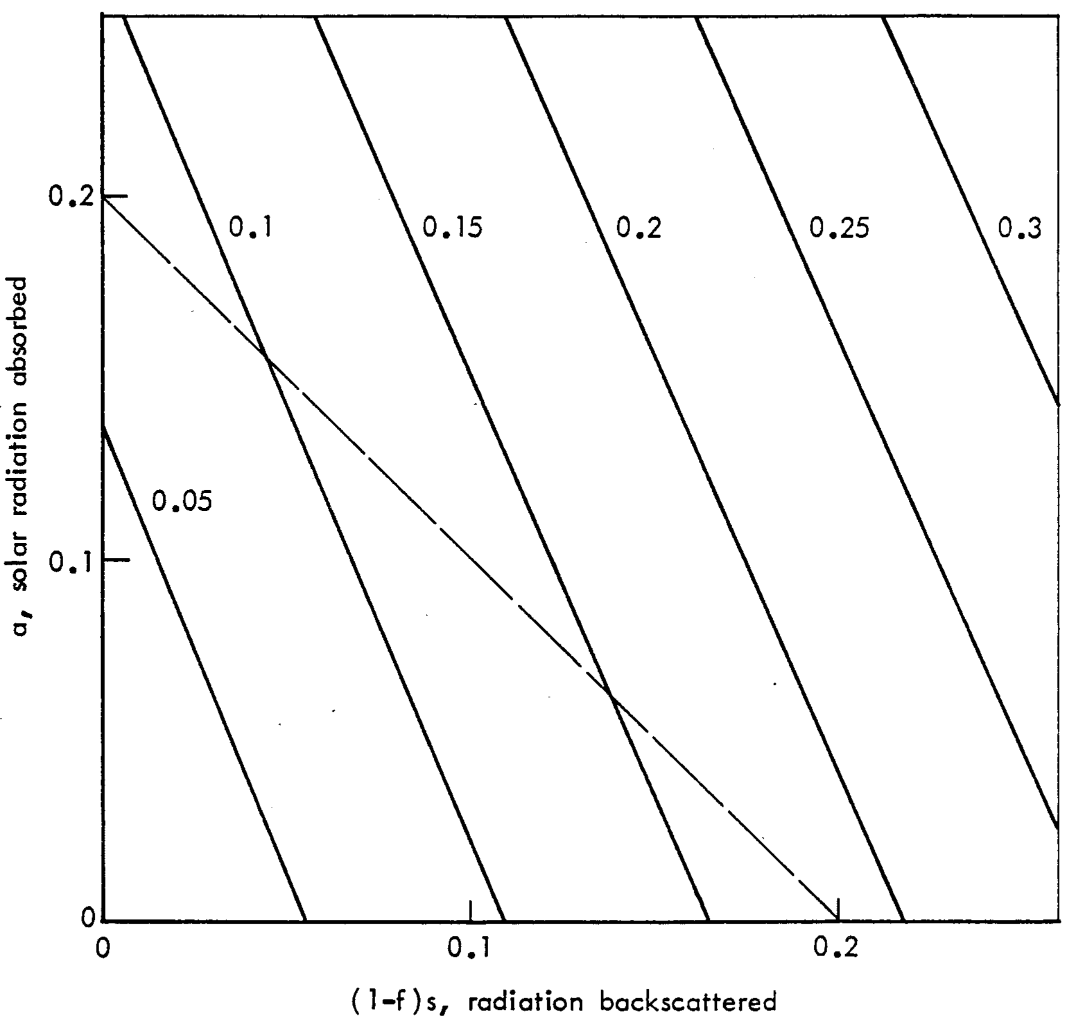

Fig. 6 — Graph of the function Φ for the earth's surface.

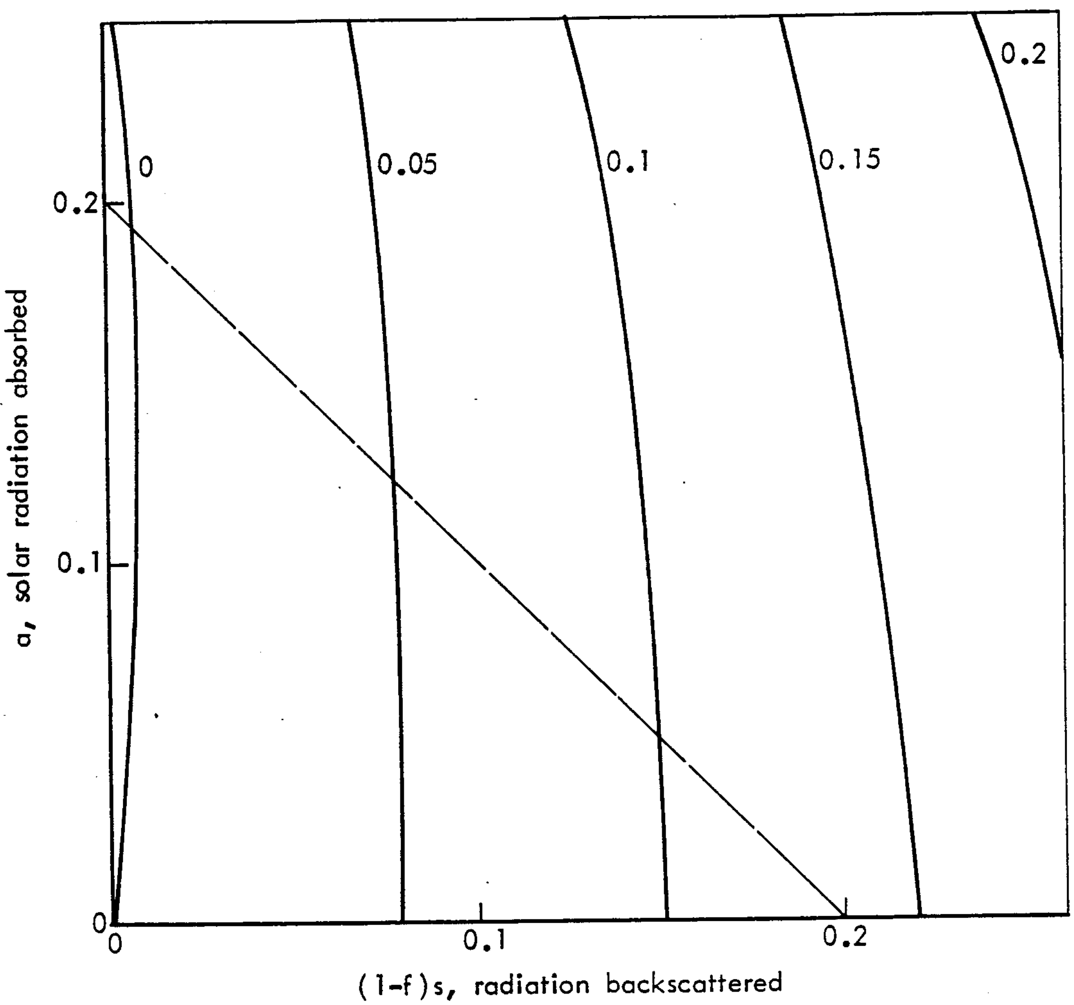

Fig. 7 — Graph of the function Φ for the atmosphere.

are interpreted in the following way. For given values -- on the vertical axis for radiation absorbed by the cloud; on the horizontal axis (1 - f)s for radiation backscattered -- the value of Φ can be determined. Then Φ is compared to the total attenuation of terrestrial radiation, and the nature of the temperature change is determined from the conditions given by Eq. (4). Consider for convenience that f is zero. Then the horizontal axis will give the total per cent scattered, s. If the total attenuation of solar radiation is 20 per cent, the dashed line between a = 0.2 and s = 0.2 will contain all possible combinations of a and s. Thus from Fig. 6 and Eq. (4), we find that if $\beta < 0.075$, only cooling is possible, but if $\beta > 0.18$, only warming is possible. For a value of β between 0.075 and 0.18 the exact values of a and s must be known in order to determine the effects. For a equal to 0.1 and s equal to 0.1, $\beta < 0.125$ would indicate cooling and $\beta > 0.125$ would indicate warming. A similar interpretation can be given to the Φ curves for the atmosphere in Fig. 7. Note, however, the different character of the curves for the atmosphere. For the same conditions discussed above (f = 0; a = 0.1; s = 0.1), we find that $\beta < 0.065$ would indicate cooling and $\beta > 0.065$ would indicate warming. Unlike the results of Humphreys, this analysis shows that the conditions for warming or cooling of the surface are not identical to those of the atmosphere. The above example shows three possibilities: the earth and atmosphere may warm, the earth and atmosphere may cool, or the earth may cool while the atmosphere warms. In Fig. 8 are given the curves of Φ for the earth (solid line) and for the atmosphere (dashed line). We see immediately that the earth and the atmosphere curves with the same value of Φ do not intersect, indicating that the earth and the atmosphere cannot both remain as they were before

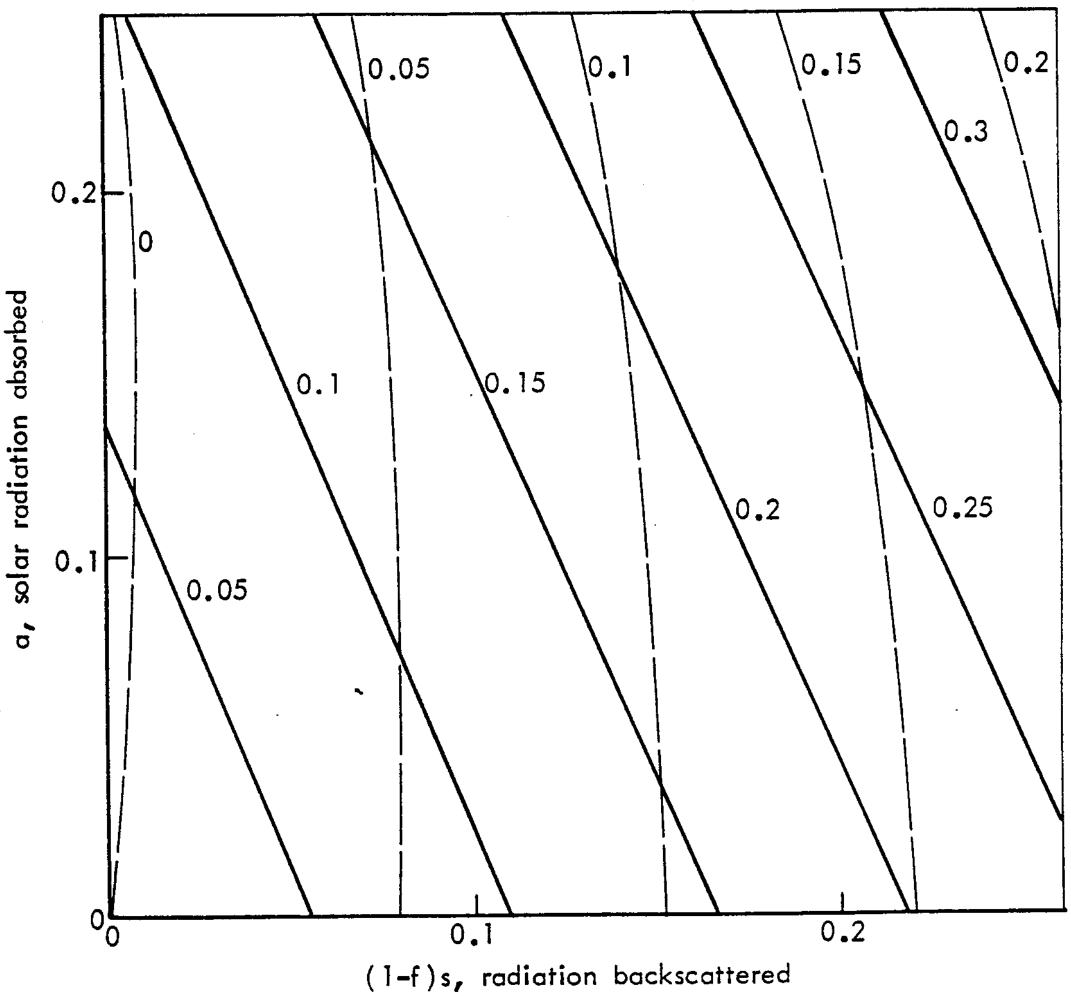

Fig. 8 — Graph of Φ_g (——) and Φ_a (— —).

the dust cloud. Therefore, by injecting a dust layer into the stratosphere, there will always be some modification of the heat balance. Only the degree to which the cloud will affect the weather-producing processes is in doubt. Another consequence of the nonintersecting Φ curves is that for any values of a and s, the Φ values for the atmosphere are always less than those for the surface. Hence in a new notation we can write

$$\begin{array}{ll} \beta < \Phi_a & \text{surface cools, atmosphere cools} \\ \Phi_a < \beta < \Phi_g & \text{surface cools, atmosphere warms} \\ \Phi_g < \beta & \text{surface warms, atmosphere warms} \end{array} \qquad (5)$$

where the first subscript, a, refers to the atmosphere and the subscript, g, refers to the surface. In addition, when $\beta < \Phi_a$ the atmosphere tends to cool less than the surface, and when $\beta > \Phi_g$ the atmosphere tends to warm more than the surface. For example, if we once again take a = 0.1, s = 0.1, and f = 0, we can compute from equations in Appendix B the values for the temperature change given in Table 2.

Table 2

	$\beta = 0.5 < \Phi_a$	$\Phi_a < \beta = .1 < \Phi_g$	$\Phi_g < \beta = .15$
Atmosphere	-1°K	+2°K	+6°K
Surface	-9°K	-2°K	+1°K

Clearly, the cloud will tend to stabilize the vertical temperature gradient.

The values given in the examples above are not intended to suggest expected changes following a nuclear war, but to serve only as an illustration of the cloud effects. Reliable estimates cannot be given without a better knowledge of the attenuating properties of the cloud. Let us, however, assume a range of reasonable values for a, s, a', and s', keeping in mind that "reasonable" does not exclude other possibilities that may be extreme. The results and assumptions for four cases are given in Table 3.

Table 3

	f	s	a	β	ΔT_g	ΔT_a
Case 1	.5	.2	0	0	-6°K	-4°K
Case 2	.75	.2	0	0	-3°K	-1.5°K
Case 3	.75	.15	.05	0	-4°K	-1.5°K
Case 4	.75	.15	.05	.05	0	+2°K

While a 20 per cent attenuation of solar radiation may be reasonably assumed just by a comparison with Krakatoa and Katmai, it is evident from the results that a knowledge of the forward scattering by the cloud and terrestrial radiation trapped by the cloud is critical in evaluating the effects of the cloud.

In computing the temperature changes presented above, no influence by the oceans was considered. The oceans may affect the temperature changes in two ways. First, because of the large heat capacity of the oceans, one might expect a lag in the temperature changes and perhaps a decrease in the magnitude of the temperature change, much as the seasonal and diurnal temperature changes are modified at land stations

with an ocean influence. Second, it follows from the first that the normal land—ocean contrasts may be modified. These contrasts play an important role in determining the mean temperature and pressure patterns of the atmosphere. It is conceivable that with even minor change over the continental areas but little or no change over the ocean areas, the distribution of pressure and temperature around the earth at a given latitude would be modified enough to produce anomalous weather patterns. The results of Arakawa, et al. (1955) seem to bear out this contention. In their studies on the causes of poor rice crops in Japan they conclude that the relationships among cool summers, bad harvests, and volcanic explosions is statistically significant. They further demonstrate that, although the magnitude and wavelengths of temperature waves along the latitude belt between 50°N and 60°N in years following eruptions do not notably differ from those in other years, the patterns differ significantly. Within this same latitude belt and over the Eurasian continent, the temperature is found to be below normal in spring, summer, and fall, but above normal in winter, a result that could be attributed to a modification of normal ocean—land contrasts.

In any theory used to predict the effects of stratospheric dust layers on the weather, proper consideration must be given to north—south variations of the heat budget since pole-to-equator temperature gradients are important in determining the nature of atmospheric disturbances. Humphreys proposed in his theory that the volcanic debris would be evenly distributed in the stratosphere. He then argues that because at high latitudes the radiation travels a greater distance through the layer, the temperatures at high latitudes should be affected

most, thus increasing the equator-to-pole temperature gradient. The critical assumption here is that the debris will be evenly distributed in the stratosphere. Experience with radioactive debris in the stratosphere tells us that the final distribution depends greatly on the altitude and the latitude of injection. Concentrations of debris injected in the equatorial stratosphere (an example similar to the eruption of Krakatoa) are found to be heaviest at low latitudes. Hence, although at high latitudes the radiation travels a greater distance through the atmosphere, the total attenuation may there be less because of the smaller concentration of debris. If we assume that the amounts of sensible and latent heat transported in the atmosphere will remain unchanged, the initial trend of the horizontal temperature gradient can be determined for a given distribution of debris in the stratosphere. For example, assume that the debris is distributed so that the net loss of solar radiation will remain constant at all latitudes. Then by using the values of f, s, a, and β in Case 1 (Table 3), one can compute the atmospheric temperature changes for 10°-latitude belts given in Table 4.

Table 4

Latitude Zone	ΔT_a (deg. C)
0 to 10°	-5.5
10 to 20°	-5.5
20 to 30°	-5
30 to 40	-4.5
40 to 50	-4
50 to 60	-3
60 to 70	-2
70 to 80	-1.5
80 to 90	-1

The horizontal temperature gradient in this example will decrease. The disagreement between this result and that of Humphreys tells us that the horizontal temperature gradient cannot be determined without first specifying the distribution of debris in the stratosphere. Debris injected at mid-latitudes, as expected in a nuclear war, could affect the radiation balance in a different way.

Finally, one must also consider the impact of the initial temperature trends on the sensible- and latent-heat transports, for the ultimate temperature distribution will depend not only on the radiational processes but also on the heat gained through the transport processes.

Thus far, only stratospheric dust layers have been considered. However, as the dust particles settle out of the stratosphere they may act as ice nuclei, enhancing the formation of cirrus clouds and thus forming another attenuating layer in the upper troposphere. This layer would not be so persistent or extensive as the stratospheric dust layer. The clouds would form only where and when the proper conditions of vertical motion and humidity prevailed. It is generally accepted, however, that ice nuclei are not abundant. One might anticipate an increase in either number of cirrus clouds or in thickness. Kuhn and Suomi (1965) have shown that thin layers of clouds in the high troposphere can significantly decrease the loss of infrared radiation to space. Sabatini and Suomi (1962) have demonstrated that the rate of atmospheric infrared cooling can be estimated by the flux of infrared radiation from the top of the atmosphere. A decrease in the flux implies a decrease in the cooling rate. Therefore, changing the number or thickness of cirrus clouds could add to the existing horizontal

variations of infrared cooling in the atmosphere. The importance of these variations has been pointed out by Suomi and Shen (1963). Exploring further, Corcoran and Horn (1965) have shown that the observed distribution of infrared cooling can provide a positive generation of potential energy in atmospheric disturbances that have short wavelengths. Thus, the growth of young storms may be partly attributed to the modification of radiational processes by the proper distribution of clouds within the disturbance. It is conceivable then, that by increasing the frequency of occurrence or thickness of cirrus clouds, the contribution of infrared cooling to the development of storms may sometimes be increased.

It may be concluded that the debris produced by a nuclear war can modify the weather. However, the nature of the changes cannot be predicted without a better understanding of the atmospheric processes involved.

Modifications Related To Large Fires

One of the consequences of nuclear detonations is fire. It is possible then, that large-scale forest fires may be started by the numerous detonations in a nuclear war. If a forest fire is extensive, there may be both short- and long-term effects on the weather.

Fire and its by-products can be classified in the same way as nuclear detonations and their by-products were in the previous section, that is, by their effects on the dynamical, hydrological, and radiational processes. For example, thunderstorms often develop in the rising air mass produced by the heat released in fire storms. Combustion products are excellent condensation nuclei and may have some effect on precipitation; smoke plumes can reduce the amount of

solar radiation. The pall of smoke from the 1950 Alberta fire was believed responsible for reducing the incoming radiation by 54 per cent and lowering the temperatures 10°F at Washington, D. C. (Wexler, 1950). These short-term effects are not expected to last for more than a few weeks. However, the destruction by fire of vegetation over a large area can produce longer-lasting climatic changes in the local burned area and perhaps beyond.

The removal of trees from forested areas can produce local climate changes sometimes so severe that regeneration may be curtailed. The expected changes in the denuded areas include hotter days and colder nights, hotter surface temperatures in summer and colder in winter, lower humidity, higher winds, deeper snow but earlier melts, and, it is thought by some, less precipitation (Zon, 1941; Hursh and Connaughton, 1938). These effects of fire on the local climate, however, may not be so important as the possible changes in the character of the surface, which can modify the heat supply to the atmosphere. The anomalous snow cover mentioned above in connection with feedback mechanisms is one example of possible weather modification resulting from changes in the character of the surface. According to D. M. Houghton (1958), the surface variations most effective in changing the energy available to the atmosphere are anomalous sea-surface temperatures, variations in the limits of frozen land and sea, anomalous snow covers in middle latitudes, and changes from moist land (or land covered by green vegetation) to desert and vice versa. Of these, changes both in the surface albedo caused by an anomalous snow cover and in the heat used for evaporation caused by the removal of vegetation may be large enough to modify the weather in an area considerably larger than the deforested

area. For example, the albedo of a "snow-covered forest" is 10 to 25 per cent, while that of "fresh snow" is 70 to 80 per cent. Thus the increase in albedo of approximately 50 per cent can reduce the supply of heat eventually available to the atmosphere by 150 cal cm^{-2} day^{-1} during late winter when the incident solar and sky radiation is near 300 cal cm^{-2} day^{-1}. We have already discussed a possible effect of an anomalous snow cover on the weather patterns.

The decrease in the amount of heat used for evaporation by removing vegetation can be estimated from data given in a study by Ziemer (1964). This study compares soil moisture in forest openings one, five, ten, and twelve years old, with the soil moisture in the surrounding forests. Ziemer found that once the surface was dry, it acted as an insulator against further loss through surface evaporation. In openings where vegetation was becoming established, more soil moisture was lost by transpiration. In the month of maximum soil-moisture depletion, the 1-year-old clearing had 6 inches more moisture per 4 feet of soil than the surrounding forest. In clearings 5, 10, and 12 years old, the excesses were 2.9, 1.2, and 0.7 inches respectively. If each of these excesses is taken as representing a loss of evaporation in summer, the heat unused for evaporation amounts to 120, 50, 20, and 10 cal cm^{-2} day^{-1} for clearings 1, 5, 10, and 12 years old respectively -- and this unused heat can be considered a source of heat previously unavailable to the atmosphere. The effects of an atmospheric heat source of this magnitude are not readily calculated. However, Namias (1962) has presented a theory for a similar situation. He suggests that synoptic conditions favorable to summer droughts in the Great Plains may be a result of the increased amounts of heat made available to the atmosphere

through decreased evaporation during warm dry spring seasons. In support of this thesis he presents the data in Table 5. Note that although a cold spring is more likely to be followed by a cold summer than by a warm one, the ratio of cold summers to warm summers increases from 101/40 to 41/11 if the springs are also wet. And the ratio of warm summers to cold summers, following a warm spring, increases from 87/57 to to 50/9 if the springs are also dry. It is reasonable to assume, then, that anomalous patterns of evaporation can have some effect on local circulation patterns.

In addition, if the results of Ziemer are representative of areas denuded by fire, the effects can be expected to continue in decreasing strengths for about five years following the burn.

The effects of anomalous sources and sinks of heat on weather patterns need further study, and certainly, the possibility of this kind of weather modification following a nuclear war should not be overlooked.

Table 5

NUMBER OF SUMMERS FOR EACH TEMPERATURE CLASS FOLLOWING NINE COMBINATIONS OF SPRING TEMPERATURE AND PRECIPITATION

(Various stations on western Great Plains)*

Spring Temperature	Spring Precipitation	Cases	Number of Summers		
			Cold	Normal	Warm
Cold	Light	60	29	21	10
	Moderate	67[sic]	31	18	19
	Heavy	83	41	31	11
	Total	...	101	70	40
Normal	Light	64	12	18	34
	Moderate	78	18	33	27
	Heavy	65	23	23	19
	Total	...	53	74	81 [sic]
Warm	Light	86	9	27	50
	Moderate	62	18	22	22
	Heavy	62	30	16	16
	Total	...	57	65	87 [sic]

*Adapted from Namias.

IV. CONCLUDING SUMMARY

The means by which a nuclear war may modify the precipitation and heat balance have been discussed. The large quantities of debris injected into the troposphere and stratosphere seem to be the most likely instrument involved in the modification. The debris can act in several ways. It may provide an additional source of nonsoluble condensation nuclei and ice nuclei. As ice nuclei it may increase the amount of thin cirrus clouds, thus affecting the radiation balance, or it may increase the efficiency of the ice-crystal mechanism of precipitation formation in super-cooled clouds. The very fine dust particles left behind in the stratosphere, if sufficient in number, will interfere with the radiation entering and leaving the atmosphere. In addition to the effects of the debris, fires ignited by nuclear detonations may, if extensive, change the surface characteristics and modify local weather patterns.

Therefore, it is possible that the weather can be modified as a result of a nuclear war. The nature, extent, and magnitude of the changes are uncertain. It must be pointed out, however, that it is not sufficient to predict the modification of the weather on the sole basis of the direct effects of a single perturbation. To understand the nature of the climatic change, one must consider both the direct and indirect effects of all relevant perturbations on all the atmospheric processes. It is this integration of the problem that requires a keener understanding of the atmosphere.

Appendix A

ESTIMATION OF THE AMOUNT OF STRATOSPHERIC DEBRIS PRODUCED BY A ONE-MT DETONATION

In some recent work, I. J. Russell* has shown that the fraction of particles of diameter D in a radioactive cloud at 20 minutes can be approximated by

$$F(D) = K\,D^{-3.25} \quad . \tag{1A}$$

This relationship holds for values of D between 1μ and 100 to 160μ (approximately the largest particles in the cloud at 20 minutes). From Eq. (1A) and estimates of the amount of material initially in the cloud and in the cloud at 20 minutes, the fraction of the cloud consisting of particles less than 2μ in diameter can be estimated. In Eq. (1A) it can be noted that F(D) approaches infinity as D approaches zero. This, of course, is not realistic, but it will be used to obtain an upper limit. A lower limit to the total mass of particles less than 2μ in diameter will be obtained by assuming a linear distribution of particles between 0 and 2μ (see Fig. 9). Thus for the upper limit, the fraction of the cloud at 20 minutes consisting of particles less than 2μ becomes

$$\frac{\int_0^2 4/3\ \pi\rho D^3\ K\ D^{-3.25}\ dD}{\int_0^{160} 4/3\ \pi\rho D^3\ K\ D^{-3.25}\ dD} = \left(\frac{2}{160}\right)^{3/4} \cong .04 \tag{2A}$$

where ρ is the density of the particles.

---
*Private communication.

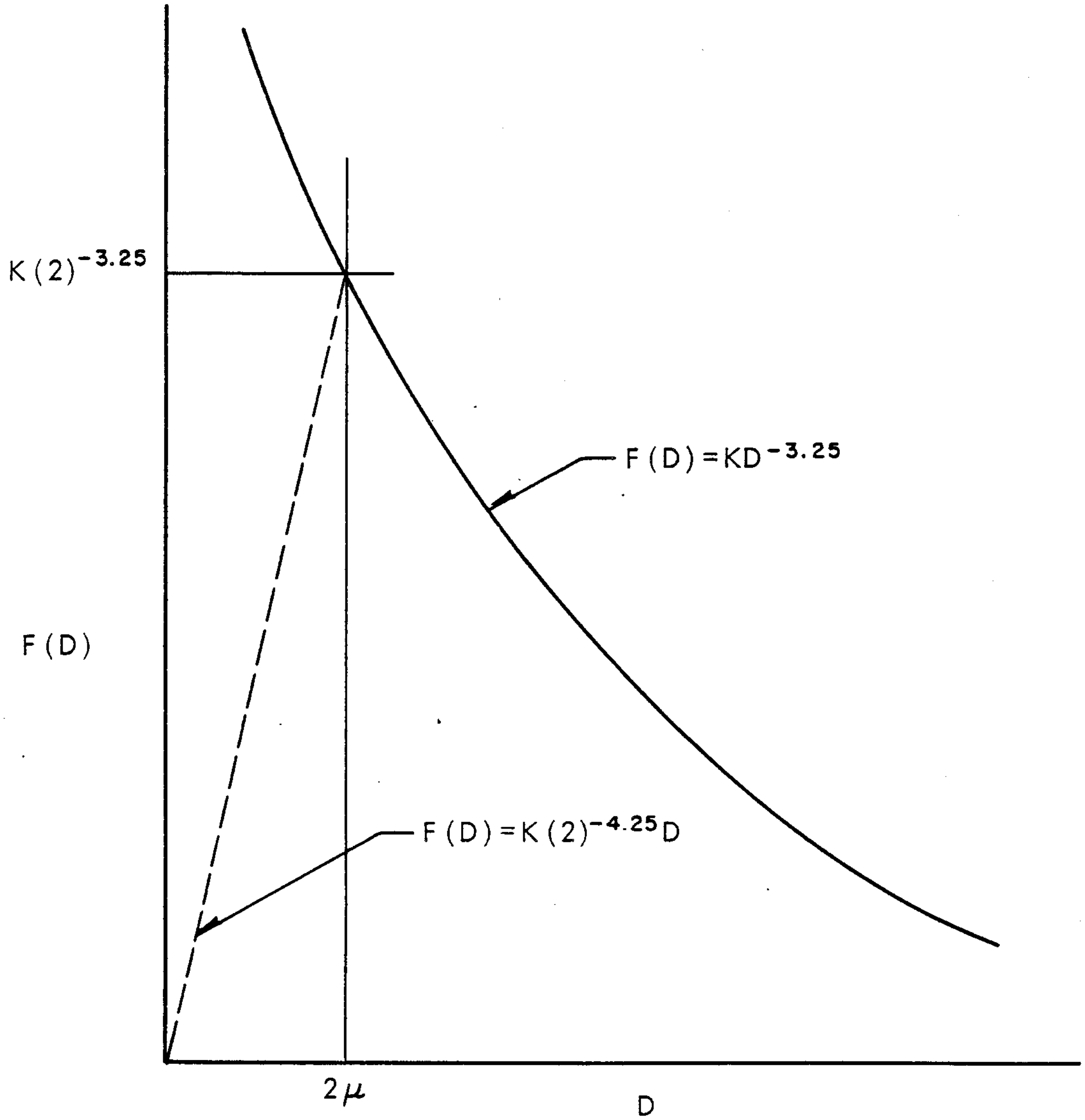

Fig. 9 — Schematic diagram of particle size distributions for upper and lower limits.

For the lower limit we find

$$\frac{\int_0^2 4/3\ \pi\rho D^3\ (2)^{-4.25} KD\ dD}{\int_0^2 4/3\ \pi\rho D^3\ (2)^{-4.25}\ KD\ dD + \int_2^{160} 4/3\ \pi\rho D^3\ K\ D^{-3.25}\ dD} = \tag{3A}$$

$$\frac{1/5\ (2)^{3/4}}{1/5\ (2)^{3/4} + 4/3\ (160)^{3/4} - 4/3\ (2)^{3/4}} \cong 0.006$$

Russell estimated that the cloud at 20 minutes contained 400 tons of material and that 1000 tons had already fallen out. Thus the percentage of the initial cloud consisting of particles less than 2μ in diameter ranges between 1.1 and 0.17 per cent.

Using one ton of material per ton of yield as an order-of-magnitude value for the amount of material put in the cloud from a surface burst, we find that between 10^3 and 10^4 tons of particles less than 2μ in diameter are produced for each megaton of yield. In terms of volume, each MT produces between 4×10^{-7} and 4×10^{-6} km^3 assuming a mean density of 2.5 tons/m^3.

Appendix B

DERIVATION OF THE EQUATIONS FOR THE EFFECTS OF A DUST CLOUD ON THE HEAT BUDGET

In deriving the equations for the effects of a dust cloud on the heat budget, it will be assumed that the mean cloudiness, the surface albedo, the atmospheric absorptivity of long- and shortwave radiation, the transmissivity of the atmosphere, the amount of evaporation, and the amount of turbulent transfer of heat from the surface will not change after the dust cloud is formed in the stratosphere. As it is likely this will not be the case, the results can be interpreted only as initial trends.

If E_{g0} is the equilibrium value of radiation emitted by the earth without a dust cloud and E_g the equilibrium value with a dust cloud, the conditions for warming or cooling, assuming the amount of heat radiated is proportional to the fourth power of the temperature, can be expressed by

$$E_g \begin{matrix} > \\ = \\ < \end{matrix} E_{g0} \quad \begin{cases} \text{warming} \\ \text{no change} \\ \text{cooling} \end{cases} \tag{1B}$$

for the earth. Similarly, for the atmosphere we have

$$E_a \begin{matrix} > \\ = \\ < \end{matrix} E_{a0} \quad \begin{cases} \text{warming} \\ \text{no change} \\ \text{cooling} \end{cases} \tag{2B}$$

For a balance of heat at the surface, the amount of radiation emitted by the surface is equal to the net gain of heat from all sources. Thus E_g is equal to the sum of the shortwave radiation absorbed

by the surface (γI), the longwave back radiation from the atmosphere (δE_a), the loss of heat by evaporation and turbulent transfer (H), and the longwave radiation from the bottom of the dust cloud that is transmitted through the atmosphere ($\tau I'$).

$$E_g = \gamma I + \delta E_a - H + \tau I' \tag{3B}$$

where I is the shortwave radiation incident on the atmosphere below the cloud, I' is the longwave radiation from the bottom of the dust cloud, δ is the fraction of the radiation absorbed in the atmosphere that is emitted toward the surface, τ is the transmissivity of the atmosphere to longwave radiation, and γ is the fraction of solar radiation absorbed by the surface.

The total amount of radiation emitted by the atmosphere is equal to the heat gained through absorption of solar radiation (αI), absorption of longwave radiation from the surface ($\alpha' E_g$), condensation and turbulent transfer (H), and absorption of longwave radiation from the dust cloud ($\alpha' I'$).

$$E_a = \alpha I + \alpha' E_g + H + \alpha' I' \tag{4B}$$

where α is the fraction of solar radiation absorbed by the atmosphere and α' (equal to $1 - \tau$) is the fraction of longwave radiation absorbed by the atmosphere.

Solving Eqs. (3B) and (4B) for E_g and E_a we get

$$E_g = I\lambda_g + I'\eta_g - H\omega_g \tag{5B}$$

$$E_a = I\lambda_a + I'\eta_a + H\omega_a \tag{6B}$$

where

$$\lambda_g = \frac{\gamma + \delta\alpha}{1 - \delta\alpha'}; \quad \eta_g = \frac{\tau + \delta\alpha'}{1 - \delta\alpha'}; \quad \omega_g = \frac{1 - \delta}{1 - \delta\alpha'}$$

$$\lambda_a = \frac{\alpha + \alpha'\gamma}{1 - \delta\alpha'}; \quad \eta_a = \frac{\alpha' + \alpha'\tau}{1 - \delta\alpha'}; \quad \omega_a = \frac{1 - \alpha'}{1 - \delta\alpha'}$$

Without a dust cloud, I' will be zero and I will be equal to the unattenuated solar radiation I_0. Thus from Inequalities (1B) and (2B) we get

$$(I - I_0)\,\lambda_g + I'\eta_g \gtreqless 0 \quad \begin{cases} \text{warming} \\ \text{no change} \\ \text{cooling} \end{cases} \tag{7B}$$

for the surface, and

$$(I - I_0)\,\lambda_a + I'\eta_a \gtreqless 0 \quad \begin{cases} \text{warming} \\ \text{no change} \\ \text{cooling} \end{cases} \tag{8B}$$

for the atmosphere.

It remains to determine the expressions for I and I' in terms of the solar radiation absorbed (a) and scattered (s) by the cloud, and the terrestrial radiation absorbed (a') and scattered (s') by the cloud.

The shortwave radiation leaving the bottom of the cloud, is equal to the sum of solar radiation transmitted through the cloud, the solar radiation scattered forward by the cloud, and the fraction of the shortwave radiation reflected by the earth and atmosphere that is scattered backward from the cloud.

$$I = I_0(1 - a - s) + fsI_0 + \rho I(1 - f)s$$

where f is the fraction of radiation scattered forward and ρ (equal to $(1 - \alpha - \gamma)$) is the albedo of the earth and atmosphere.

Solving for I

$$I = I_0 \left[\frac{1 - a - (1 - f)s}{1 - \rho\ (1 - f)s} \right] = I_0 M \qquad (9B)$$

If it is assumed that the longwave radiation from the cloud is emitted and scattered equally in the upward and downward directions, we find

$$I' = \frac{a}{2} (I_0 + \rho I) + \frac{1}{2} (a' + s')(\tau E_g + \mu E_a) \qquad (10B)$$

where μ (equal to $1 - \delta$) is the fraction of radiation emitted upward by the atmosphere, $a(I_0 + \rho I)$ is the amount of shortwave radiation absorbed by the cloud, and $(a' + s')(\tau E_g + \mu E_a)$ is the longwave radiation reaching the bottom of the cloud and absorbed or scattered by it. Adding Eqs. (3B) and (4B) and recalling that $\tau = 1 - \alpha'$ and $\mu = 1 - \delta$, we find

$$(\tau E_g + \mu E_a) = (\gamma + \alpha)I + I'$$

Substituting this relation into Eq. (10B), replacing I with $I_0 M$ from Eq. (9B), and solving for I', we get

$$I' = I_0 \left[\frac{\frac{1}{2} (a' + s')M(\gamma + \alpha) + \frac{a}{2} (1 + \rho M)}{1 - \frac{1}{2} (a' + s')} \right] \qquad (11B)$$

Thus the equilibrium value of the radiation emitted by the earth and atmosphere when a dust cloud is present can be determined using Eqs. (9B) and (11B) with (5B) for the earth, (6B) for the atmosphere, and by any set of values for a, s, a', and s'.

A further development will yield the conditions for warming or cooling in terms of the terrestrial radiation trapped by the dust cloud. Substituting Eqs. (9B) and (10B) into either Eq. (7B) for the earth or Eq. (8B) for the atmosphere we get, on solving for $\frac{(a' + s')}{2}$

$$\frac{(a' + s')}{2} \gtreqless \left[\frac{(1 - M)\lambda - \frac{a}{2}(1 + M\rho)\eta}{(1 - M)\lambda + M(1 - \rho)\eta} \right] \begin{cases} \text{warming} \\ \text{no change} \\ \text{cooling} \end{cases} \tag{12B}$$

where λ and η become λ_g and η_g if Eq. (7B) is used or λ_a and η_a if Eq. (8B) is used.

For convenience, we may replace $\frac{(a' + s')}{2}$ with β, and the right side of Eq. (12B) with $\Phi(a, s, f, \rho, \gamma, \alpha, \tau, \delta)$. Thus

$$\beta \gtreqless \Phi(a, s, f, \rho, \gamma, \alpha, \tau, \delta) \begin{cases} \text{warming} \\ \text{no change} \\ \text{cooling} \end{cases} \tag{13B}$$

The values of ρ, γ, α, τ, and δ can be obtained from radiation budget studies. The values used here are those given by London (1957). Thus for assumed values of a, s, and f, the value of Φ can be computed and compared with the amount of terrestrial radiation trapped by the cloud (β) to determine the nature of the temperature changes.

Values of Φ for the surface and the atmosphere are given for all values of a, s, and f in Figs. 10 and 11.

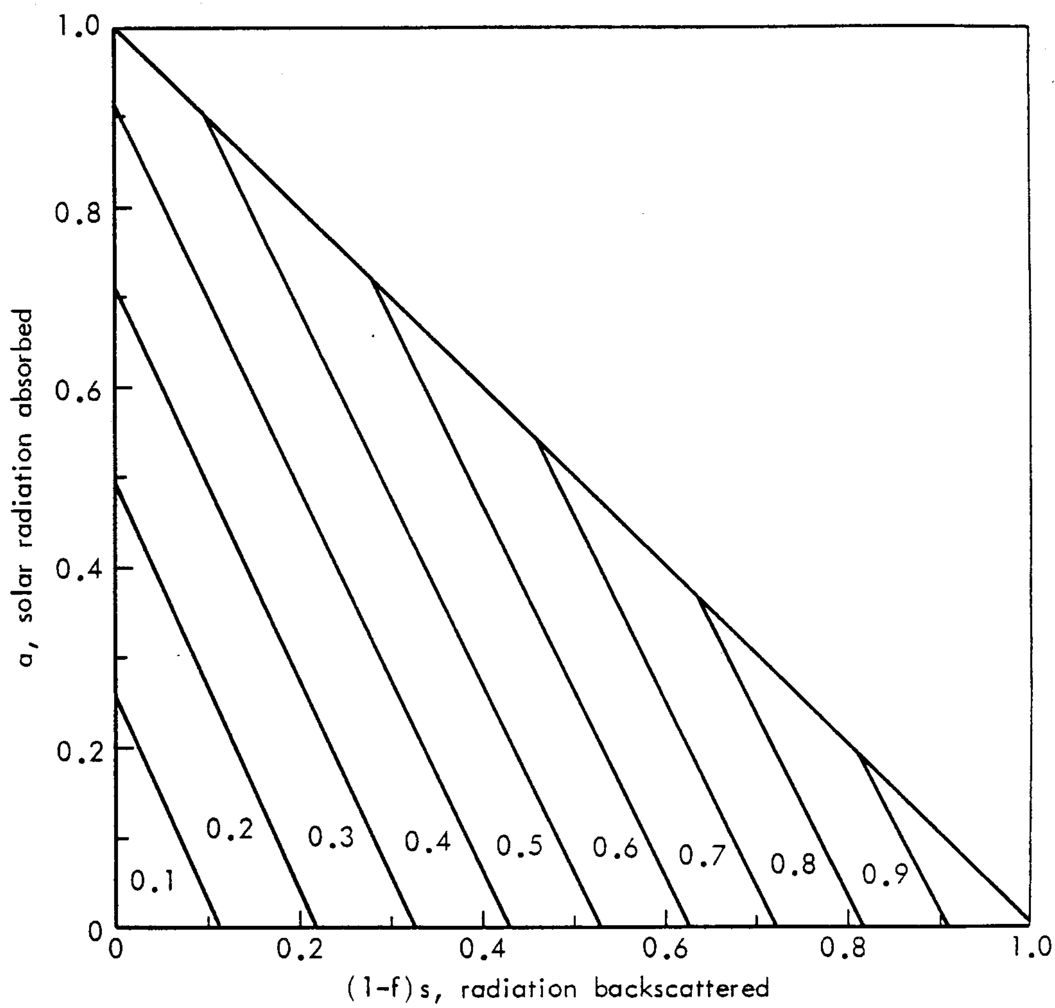

Fig. 10 — Graph of the function Φ for the earth's surface.

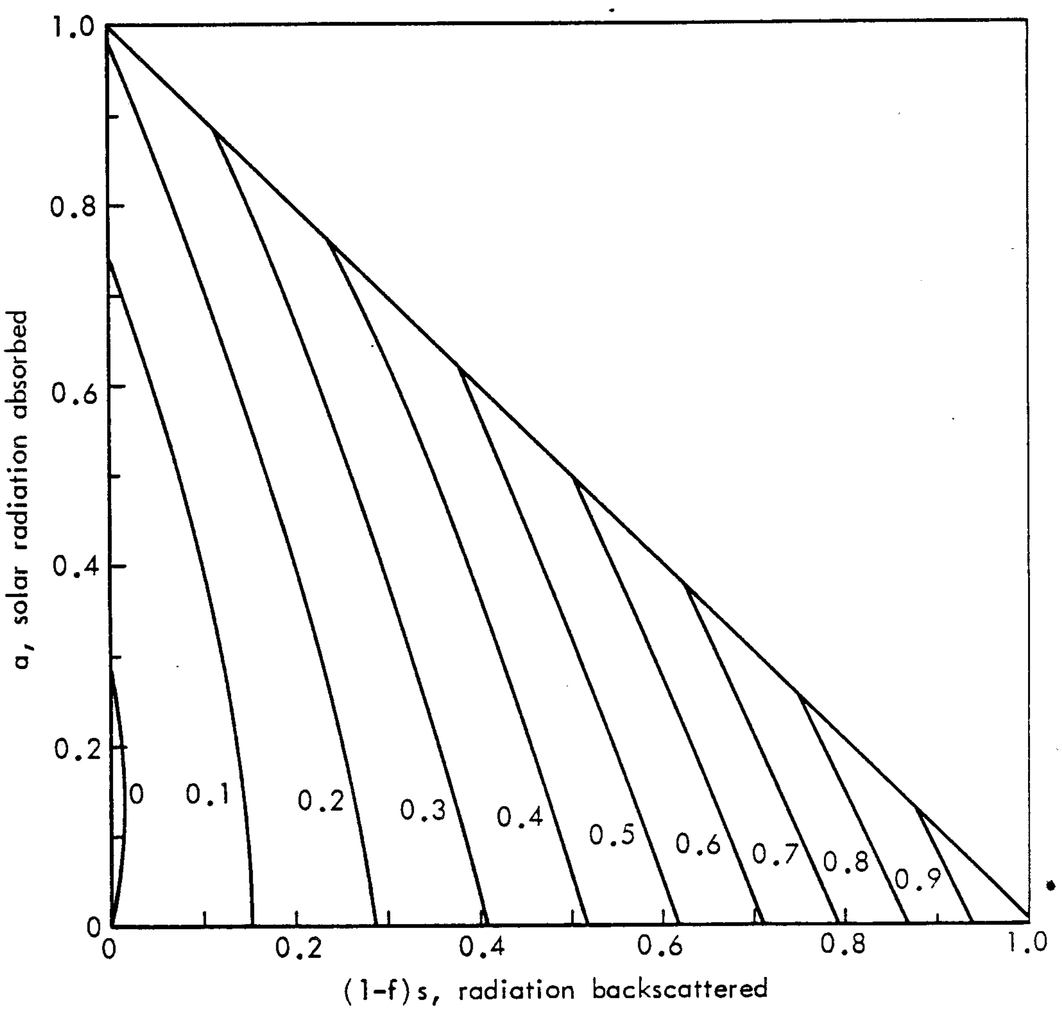

Fig. 11 — Graph of the function Φ for the atmosphere.

BIBLIOGRAPHY

Arakawa, A., T. Fujita, H. Itoo, Y. Masuda, S. Matsumoto, T. Murakami, T. Ozawa, E. Suzuki, M. Takeuchi, and K. Tomatsu, "Climate Abnormalities as Related to the Explosions of Volcano and Hydrogen Bomb," Geophys. Mag. (Tokyo), 26, 1955, pp. 231-255.

Arakawa, H., "Possible Atmospheric Disturbances and Damages to the Rice Crops in Northern Japan that May be Caused by Experimentation with Nuclear Weapons," Geophys. Mag. (Tokyo), 26, 1955, pp. 125-134.

Arakawa, H., and K. Tsutsumi, "A Decrease in the Normal Incidence Radiation Values for 1953 and 1954 and its Possible Cause," Geophys. Mag. (Tokyo), 27, 1956, pp. 205-208.

Ayres, R. U., "Effects of Thermonuclear War on Weather and Climate," Special Aspects of Environment Resulting from Various Kinds of Nuclear Wars, Part III, Annex IV, Hudson Institute, Inc., HI-303-RR/A IV, January 1964.

Bengelsdorf, I., "Can the Atom Change the Weather?" Saturday Rev., 7 July 1956, pp. 31-37.

Brooks, C. E. P., and T. M. Hunt, "The Influence of Explosive Volcanic Eruptions on the Subsequent Pressure Distribution over Western Europe," Meteorol. Mag., 64, November 1929, pp. 226-229.

Budyko, M. I., The Heat Balance of the Earth's Surface, trans. by Nina A. Stepanova, U.S. Weather Bureau, 1958, 259 pp., from "Teplovoi Balans Zemnoi Poverkhnosti," Gidrometeorologischeskoe Izdates'stvo, Leningrad, 1956, 255 pp.

Corcoran, J. L., and L. H. Horn, "The Role of Synoptic Scale Variations of Infrared Radiation in the Generation of Available Potential Energy," J. Geophys. Res., 70, No. 18, 15 September 1965, pp. 4521-4527.

Davis, P. A., "An Analysis of the Atmospheric Heat Budget," J. Atmospheric Sci., 20, No. 1, January 1963, pp. 5-22.

Dickson, R. R., "A Synoptic Characterization of the Thermal Nature of the Earth's Surface," Monthly Weather Rev., 92, No. 5, May 1964, pp. 195-201.

Enger, L., and S. Fritz, "Opacity of the Sky after July 1953," Monthly Weather Rev., 84, No. 8, August 1956, pp. 301-304.

Federov, E. K., "The Influence of Atomic Explosions on Meteorological Processes," trans. by R. H. Ponting, J. Nucl. Energy II, 5, 1957, pp. 135-145, from Atomnaya Energiya, 1, No. 5, 1956, p. 103.

Flowers, E. C., and H. J. Viebrook, "Solar Radiation: An Anomalous Decrease of Direct Solar Radiation," Science, 148, No. 3669, 23 April 1965, pp. 493-494.

Friend, J. P. (ed.), "The Application of HASP Data," DASA 1300, The High-Altitude Sampling Program, 4, 31 August 1961, pp. 54-76.

Greenfield, S. M., Ionization of Radioactive Particles in the Free Air, The RAND Corporation, RM-1153-AEC, July 1955.

Holzman, B., "The Effect of Atomic Bomb Explosions on Weather," Weatherwise, 4, No. 1, February 1951, pp. 3-4, 9.

Holzman, B., "Have Atom Bomb Tests Fouled up the Weather?" Look, 11 August 1953, pp. 32-33.

Houghton, D. M., "Heat Sources and Sinks at the Earth's Surface," Meteorol. Mag., 87, May 1958, pp. 132-143.

Houghton, H. G., "On the Annual Heat Balance of the Northern Hemisphere," J. Meteorol., 11, No. 1, February 1954, pp. 1-9.

Humphreys, W. J., "Volcanic Dust and Other Factors in the Production of Climatic Changes and Their Possible Relation to Ice Ages," J. Franklin Inst., 176, No. 2, August 1913, pp. 131-172.

Humphreys, H. J., "Factors of Climatic Control," Physics of the Air, Part V, Third Edition, Second Impression, McGraw-Hill Book Company, 1940, pp. 577-635.

Hursh, C. R., and C. A. Connaughton, "Effects of Forests Upon Local Climate," J. Forestry, 36, 1938, pp. 864-866.

Kimball, H. H., "Solar Radiation as a Meteorological Factor," Monthly Weather Rev., 59, December 1931, pp. 472-479.

Kuhn, P. M., and V. E. Suomi, "Airborne Radiometer Measurements of Effects of Particulates on Terrestrial Flux," J. Appl. Meteorol., 4, No. 2, April 1965, pp. 246-252.

London, J., A Study of the Atmospheric Heat Balance, Final Report, Contract AF 19(122)-165, AFCRC-TR, 57-267, ASTIA No. 117227, Research Division, College of Engineering, New York University, July 1957.

Machta, L., and D. L. Harris, "Effects of Atomic Explosions on Weather," Science, 121, 21 January 1955, pp. 75-81.

Mason, B. J., "Atomic Explosions and the Weather," Weather, 10, No. 5, May 1955, pp. 138-141.

Mason, B. J., "Man's Influence on the Weather," Advancement of Science, 12, No. 49, London, June 1956, pp. 498-504.

Mason, B. J., "Ice-Nucleating Properties of Clay Minerals and Stony Meteorites," Quart. J. Roy. Meteorol. Soc., 86, No. 370, October 1960, pp. 552-556.

Mason, B. J., Clouds, Rain, and Rainmaking, Cambridge University Press, 1962.

McDonald, J. E., "Cloud Nucleation on Insoluble Particles," J. Atmospheric Sci., 21, No. 1, January 1964, pp. 109-116.

Mitchell, J. M., Jr., "Recent Secular Changes of Global Temperature," Solar Variations, Climatic Change, and Related Geophysical Problems, Ann. N.Y. Acad. Sci., 95, R. W. Fairbridge (Conference ed.), and F. N. Furness (ed.), 5 October 1961.

Namias, J., "Influences of Abnormal Surface Heat Sources and Sinks on Atmospheric Behavior," Proceedings of the International Symposium on Numerical Weather Prediction in Tokyo, November 7-13, 1960, Sigekaia Syono (ed.), J. Meteorol. Soc. Japan, March 1962, pp. 615-627.

Ohring, G., and J. Mariano, "Changes in the Amount of Cloudiness and the Average Surface Temperature of the Earth," J. Atmospheric Sci., 21, No. 4, July 1964, pp. 448-450.

Peixoto, J. P., "On the Global Water Vapor Balance and the Hydrological Cycle," Studies of the Atmospheric General Circulation, IV, MIT Department of Meteorology, Final Report, Planetary Circulations Project, 30 April 1963, pp. 198-209.

Rakipova, L. R., "Influence of Atomic Explosions on the Atmosphere," Meteorology and Hydrology, No. 6, June 1957, Leningrad, pp. 53-57.

Sabatini, R. R., and V. E. Suomi, "On the Possibility of Atmospheric Infrared Cooling Estimates from Satellite Observations," J. Atmospheric Sci., 19, No. 4, July 1962, pp. 349-350.

Shaw, P. J. R., "Nuclear Explosions and the Weather," Australian Meteorol. Mag., No. 36, March 1962, pp. 39-40.

Smagorinsky, J., "The Dynamical Influence of Large-Scale Heat Sources and Sinks on the Quasi-Stationary Mean Motions of the Atmosphere," Quart. J. Roy. Meteorol. Soc., 79, No. 341, July 1953, pp. 342-366.

Staley, D. O., "Evacuation of Potential-Vorticity Changes Near the Tropopause and the Related Vertical Motions, Vertical Advection of Vorticity, and Transfer of Radioactive Debris from Stratosphere to Troposphere," J. Meteorol., 17, No. 6, December 1960, pp. 591-620.

Starr, V. P., and J. P. Peixoto, "On the Eddy Flux of Water Vapor in the Northern Hemisphere," Studies of the Atmospheric General Circulation, IV, MIT Department of Meteorology, Final Report, Planetary Circulations Project, 30 April 1963, pp. 162-197.

Starr, V. P., J. P. Peixoto, and G. C. Livadas, "On the Meridional Flux of Water Vapor in the Northern Hemisphere," Studies of the Atmospheric General Circulation, II, MIT Department of Meteorology, General Circulation Project, AFCRC-TR-58-204, 31 December 1957, pp. 124-143.

Starr, V. P., and R. M. White, "Direct Measurement of the Hemispheric Poleward Flux of Water Vapor," Studies of the Atmospheric General Circulation Project, II, MIT Department of Meteorology, General Circulation Project, AFCRC-TR-58-204, 31 December 1957, pp. 100-108.

Stonier, T., Nuclear Disaster, Meridian Books (M 154), The World Publishing Company, Cleveland and New York, December 1963.

Suomi, V. E., and W. C. Shen, "Horizontal Variation of Infrared Cooling and the Generation of Eddy Available Potential Energy," J. Atmospheric Sci., 20, No. 1, January 1963, pp. 62-65.

Sutcliffe, R. C., "Water Balance and the General Circulation of the Atmosphere," Quart. J. Roy. Meteorol. Soc., 82, No. 354, October 1956, pp. 385-395.

Sutton, G., "Thermonuclear Explosions and the Weather," Nature, 175, No. 4451, 19 February 1955, pp. 319-321.

Van de Hulst, H. C., Light Scattering by Small Particles, John Wiley and Sons, Inc., 1957, pp. 131-171.

Wexler, H., "The Great Smoke Pall--September 24-30, 1950," Weatherwise, 3, No. 6, December 1950, pp. 129-134, 142.

Wexler, H., "On the Effects of Volcanic Dust on Insolation and Weather (I)," Bull. Am. Meteorol. Soc., 32, No. 1, January 1951, pp. 10-15.

Wexler, H., "Spread of the Krakatoa Volcanic Dust Cloud as Related to High-Level Circulation," Bull. Am. Meteorol. Soc., 32, No. 2, February 1951, pp. 48-51.

Wexler, H., "Volcanoes and World Climate," Sci. Am., 186, No. 4, April 1952, pp. 74-80.

Wexler, H., "Radiation Balance of the Earth as a Factor in Climatic Change," Chapter 5, Climatic Change, Harvard University Press, Cambridge, 1953, H. Shapley (ed.), Library of Congress # 53-9041.

Wexler, H., and W. A. Hass, "Global Atmospheric Pressure Effects of the October 30, 1961 Explosion," J. Geophys. Res., 67, No. 10, September 1962, pp. 3875-3887.

Wolff, E. L. (ed.), "Aerosols--Their Complex Role in Rainfall," NCAR Quart., No. 9, January 1965.

Ziemer, R. R., "Summer Evapotranspiration Trends as Related to Time After Logging Forests in Sierra Nevada," J. Geophys. Res., 69, No. 4, 15 February 1964, pp. 615-620.

Zon, R., "Climate and the Nation's Forest," 1941 Year Book of Agriculture, Climate and Man, United States Department of Agriculture, pp. 477-498.